Frank Weber Hughes

SALT FLUX AND MIXING PROCESSES IN THE COLUMBIA RIVER ESTUARY DURING HIGH DISCHARGE.

SALT FLUX AND MIXING PROCESSES IN THE COLUMBIA RIVER ESTUARY DURING HIGH DISCHARGE

by

FRANK WEBER HUGHES

A thesis submitted in partial fulfillment
of the requirements for the degree of

MASTER OF SCIENCE

UNIVERSITY OF WASHINGTON

1968

In presenting this thesis in partial fulfillment of the requirements for an advanced degree at the University of Washington I agree that the Library shall make it freely available for inspection. I further agree that permission for extensive copying of this thesis for scholarly purposes may be granted by my major professor, or, in his absence, by the Director of Libraries. It is understood that any copying or publication of this thesis for financial gain shall not be allowed without my written permission.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

ACKNOWLEDGEMENTS

The author wishes to express his gratitude to those members of the faculty and staff of the Department of Oceanography who offered criticisms and suggestions in the preparation of this thesis. He is especially indebted to Dr. Donald Hansen for his continuing encouragement and advice during the course of this investigation.

He is grateful to Dr. Maurice Rattray, Jr., Dr. Karl Banse, and Dr. Donald Hansen for their critical reading and correction of the text and figures of this thesis; to Dr. Clifford Barnes whose suggestions and guidance led to this research; to Juraj Dworski and Paul Lu for their aid in preparation of the computer programs; to Donald Doyle and his staff for drafting services; to Robert Robe who assisted in the calculations; to Knut Aagaard for his editorial comments; and to Leilonie Gillespie for typing services.

This research was performed while the author was on active duty with the United States Navy as a participant in the Junior Line Officers Advanced Scientific Education Program (Burke Program) administered by the Bureau of Naval Personnel. Financial support for the research came in part from AEC Contract AT(45-1)-1725.

ABSTRACT

The Columbia River Estuary is shown to be both qualitatively and quantitatively different from other estuaries for which conclusive studies of the salt balance have been made. The estuary is considered to be that part of the river influenced by the intrusion of sea salt. It is characterized by an unusual combination of strong tidal currents and great river discharge. Despite large horizontal and vertical salinity gradients, the density-current structure of the estuary is weak. Analysis of the processes maintaining the salt balance during the high discharge stage indicates that the salt advected seaward with the river flow is balanced primarily by an upstream flux of salt associated with covariance of tidal-period fluctuations in current and salinity and by advective shear effect. A stronger flood current is found on the north side of the estuary coupled with stronger ebb conditions on the south side. Coriolis effect is shown to have at times a lesser effect on the dynamics of the estuary than do accelerations related to the shape of the channel. Effective values of the coefficients of horizontal and vertical eddy diffusivity are estimated and compared with similar values estimated for other estuaries by other workers.

1. INTRODUCTION

1.1 Scope and Objectives

An estuary may be defined as a semi-enclosed coastal body of water having free communication with the open sea and an influx of fresh water, deriving from land drainage, which measurably dilutes the sea water. The water at any point within the estuary consists of a mixture of fresh water and sea water with the proportions, and hence the density of the water, varying with position. Estuaries derive many of their features from this internal density distribution; for this reason, the Columbia River Estuary will be considered to be the part of the river influenced by the intrusion of sea salts.

The density distribution gives rise to pressure gradients which tend to produce a circulation in which water moves seaward in an upper layer and upstream (or landward) in a lower layer (circulation refers to the mean flow over one or more tidal cycles). Estuarine circulation includes this convective motion and the net motion which is necessary to remove the excess of fresh water that is continually being added. The circulation may also be influenced by the tide, gravitational stability, and wind. Gravitational stability tends to reduce the convective circulation. Tidal currents, however, may generate turbulence, promoting vertical mixing and thus reducing the stability.

The present study attempts to analyze the salinity and velocity distributions found in the Columbia River Estuary during a period of high

discharge. The objectives are to define the circulation, to compute the salt flux at points in the estuary, and to estimate the relative importance of the processes contributing to the salt balance. Effective values of the coefficients of horizontal and vertical eddy diffusivity will be estimated so that the estuary can be compared with others for which similar estimates have been made.

1.2 Historical Perspective

Many of the early studies of estuarine circulation were concerned with the rates at which dissolved and suspended pollutants were removed from the estuary by tidal motions. The processes which bring about the dispersion and removal of the dissolved and suspended pollutants that are discharged into an estuary are identical to those governing the mixing of fresh water and sea water. Studies which deal with the rate at which the fraction of fresh water or the concentration of water-borne substances decreases within the estuary are called flushing studies.

The earliest attempts to estimate the magnitude of tidal flushing effects made use of the classical tidal prism method which assumes complete mixing with the freshwater of the sea water entering on the flood tide and complete removal of the mixed water on the ebb tide such that a fresh supply of sea water is available for the next flood tide. This method is based on a physical process which does not occur. The tidal prism has been defined by Ketchum (1951) as the difference between the volumes of water in the estuary at high and low tides which implies that part of the high water volume is contributed by the river flow. He proposed a modification of the tidal prism method in which the estuary was divided into segments based upon the tidal excursion. Arons and Stommel (1951) expanded his concept into a mixing length theory in which the mixing along the estuary was

characterized by an eddy diffusivity proportional to the product of the length of the tidal excursion and the amplitude of the tidal current.

Since the above methods did not appear to be sufficiently accurate to be generally applicable, Stommel (1953) proposed a semi-empirical technique to determine the horizontal diffusivity in sectionally homogeneous estuaries. Bowden (1963) believed that the method of Stommel would be applicable to estuaries in which the salinity and the velocity vary over a cross section, provided that one deals only with the mean values over the section and neglects the details of the processes occurring. This approach, however, suppresses knowledge of vertical variations and considers upstream salt advection as one-dimensional diffusion. Bowden, therefore, termed the coefficient obtained an effective coefficient of horizontal eddy diffusion. Bowden also gave a method for estimating the effective values of the vertical eddy diffusivity, K_z, corresponding to the observed distribution of salinity in the River Mersey. To do so, however, he made the assumption that horizontal advection and vertical eddy diffusion are the dominant processes controlling the salt distribution as was found for the James River (Pritchard, 1954).

Pritchard (1954) investigated in detail the salt balance of the James River, a coastal plain estuary. A study of the salt balance in a fjord estuary was made by McAlister, Rattray, and Barnes (1959).

Until very recently, the majority of studies of the Columbia River Estuary have been concerned with the phenomena affecting the improvement and maintenance of the river and its entrance for navigation. Results of these studies were summarized by Lockett (1963), who stated that energy dissipated by hydraulic forces in the entrance and estuary area exceeds that dissipated in any similarly improved area in the world and "gives the

Columbia Estuary and Entrance a uniqueness beyond compare" (Lockett, 1963, p. 750).

The results of Hansen and Rattray (1965) implied that the Columbia River Estuary is qualitatively different from other estuaries for which detailed studies of the salt balance have been made. Hansen and Rattray compared estimated eddy coefficients for the Columbia at both high and low discharge stages with those estimated for other estuaries by Bowden (1963) and Hansen (1964). Their results implied that the maintenance of the longitudinal salt balance in the Columbia is different from that generally accepted for stratified estuaries; it is not maintained primarily through the advection of salt by the nontidal circulation. This result is in accordance with the relatively weak density-current structure found in the estuary despite its marked salinity stratification.

The influence of the horizontal density gradient on the velocity distribution in the Columbia River Estuary was examined by O'Brien (1952). His method of separating the effects of tidal currents, density currents, and freshwater discharge was only approximate and did not consider the role of the currents in maintaining the salinity distribution. He did, however, estimate the magnitudes of the different currents. Using data taken during a period of high discharge, he concluded that in the Columbia River Estuary the fresh water flow velocity is always greater than the velocity of the density current.

Hansen (1965) studied currents and mixing in the estuary of the Columbia during a period of low discharge. He achieved a more nearly complete separation of the advective, tidal, and turbulent mixing processes controlling the longitudinal flux of materials in an estuary than did Pritchard (1954) and Bowden (1963).

Examination of the eddy coefficients estimated by Hansen and Rattray (1965) for high and low discharge stages of the Columbia shows that the period of high discharge appears to differ not only from that of other estuaries which have been studied but also from the low discharge period in the Columbia. The difference is found in the processes which dominate the longitudinal and vertical fluxes of materials in the estuary. In an attempt to understand the processes controlling the salt distribution in the Columbia River Estuary, the present study examines the salt flux during a period of high discharge using an adaptation of the approach outlined by Hansen (1965).

The probable pollutant distribution and flushing times for the Columbia River Estuary were calculated by Neal (1965) for the low and intermediate discharge stages. Flushing time was defined as the average time required for river water with its contained pollution to move through the surveyed area. Neal's computations of flushing times were based on the modified tidal prism method (Ketchum, 1951), the fraction of fresh water method (Ketchum, 1955), and the diffusion equation method (Stommel, 1953). He found the estuary to have a relatively short flushing time even under low discharge conditions. His computations show that combinations of tidal flow and freshwater discharge may vary the flushing time from nearly ten tidal cycles to as few as two tidal cycles. These are short flushing times compared to those calculated for other estuaries. Neal also showed that north-south variations in the flow produce pollutant distributions not normally expected in an estuary. On the basis of the lateral variations in flow, he treated the estuary as consisting of two separate channels. This method does not appear to have been previously applied to an investigation of the lateral variations of flow in an estuary.

1.3 The Columbia River Estuary

1.3.1 Description

The Columbia River discharges into the Pacific Ocean between Cape Disappointment, a high rocky headland, on the north and Clatsop Spit, a low sand point, on the south (Fig. 1). The river mouth is taken to be an imaginary line connecting the seaward ends of the two jetties. The Columbia River Estuary will be considered to be that part of the river influenced by the intrusion of sea salts rather than that part of the river which is subject to tidal influence; the effect of the Pacific Ocean tides can be measured 234 km upstream at the Bonneville Dam where the diurnal range is 0.2 m. The tide at the mouth of the Columbia River is of the mixed semidiurnal type with a diurnal inequality. The diurnal range is 2.3 m (7.5 ft) and the extreme range greater than 4 m (14 ft) (Lockett, 1963; Hansen, 1965). Within 40 km of the mouth the current at all depths is reversed on flood tide during all seasons of the year. Current reversal has been measured up to 84 km from the mouth during the low discharge period (Budinger, Coachman, and Barnes, 1964; Hansen, 1965).

A concise description of the lower Columbia River, including the estuary, is given by the U. S. Army Corps of Engineers (1960, Vol. I). The river narrows and widens intermittently from approximately 14.5 km (9 miles) wide at a distance 32.2 km (20 miles) from the mouth to 3.2 km (2 miles) wide between the entrance jetties. The portion of the estuary between Point Adams and Harrington Point consists of winding channels separated by islands, bars, and shoals, the exposure of which depends on the flow conditions and the stage of the tide. Several large bays, as well as prominent points and headlands, occur along both the north and south sides

of the river. Two main river channels exist from Sand Island upstream to Tongue Point with the ship channel on the south side. The Corps of Engineers stated that because of its longer route the southern channel carries less flow than the northern channel. Between these two channels are numerous, small connecting channels.

The flow of water in the Columbia River is subject to large, seasonal variations. Maximum discharge occurs from late May through early July and corresponds to the period of maximum snow melt in the headwaters while minimum discharge occurs from late August through March. The mean annual discharge of the Columbia River at its mouth has been estimated at 7300 m^3/s (Budinger, Coachman, and Barnes, 1964). The minimum freshwater discharge has been estimated at 1850 m^3/s, while the maximum recorded discharge was about 34000 m^3/s (Hickson and Rodolf, 1951). A more nearly representative range of discharges is from 3000 m^3/s to approximately 20000 m^3/s (Hansen, 1965). Estimates of the tidal prism range from 0.58 km^3 (Lockett, 1963) to 0.85 km^3 (Budinger, Coachman, and Barnes, 1964). This volume, by definition, depends upon the range of the tides at the time of the estimate. Based on the diurnal range of the tide, 2.3 m, the tidal prism is 0.75 km^3. Using this estimate, Hansen (1965) computed that, although the tides are moderately strong, the flow ratio, which is the ratio of river discharge during a 12.4 hour tidal cycle to the tidal prism, ranges from 0.2 to 1.0, implying that the discharge current is 0.1 to 0.5 the tidal current at the mouth.

More complete descriptions of the hydrology of the Columbia River and of the estuary and entrance are given by Lockett (1963) and by Budinger, Coachman, and Barnes (1964).

1.3.2 Salinity Intrusion

The salinity intrusion depends upon both the river discharge and the tidal stage. Data compiled by Oregon State University (Burt, 1956; Burt and McAlister, 1958) and the U. S. Army Corps of Engineers (1960) indicate that the intrusion rarely extends past the Harrington Point section (Fig. 1). At high tidal stages both during periods of high and intermediate discharge, measurable salinity is found as far east as the section between Astoria and Cliff Point, approximately 24 km from the mouth. During low tide and high river discharge, the maximum salinity intrusion does not reach the Clatsop Spit section and so is limited to the western 8 km of the estuary.

1.3.3 Classification of the Estuary

There are a number of methods for classification of estuaries. A method based upon salinity structure was given by Stommel and Farmer (1952). Pritchard (1952) initially classified estuaries in terms of the relationship between fresh water inflow and evaporation and on the basis of geomorphology. In Pritchard's system, the Columbia River Estuary is a positive coastal plain estuary.

Coastal plain estuaries have been further subdivided by Pritchard (1955) according to the processes which control the salt balance. He described a sequence of estuarine types ranging from a highly stratified, salt-wedge estuary to a well-mixed, homogeneous estuary. According to Pritchard, estuaries shift from one type to another with variations in river discharge, tidal velocities, width, and depth.

Burt and McAlister (1959) placed the estuaries in Oregon in one of three of the four classifications given by Pritchard (1955): type A, the

salt-wedge; type B, partially-mixed; or type D, well-mixed. They listed the flow ratio as one of the factors that determines the mixing in an estuary and gave numerical values for flow ratios corresponding to each type of estuary found in Oregon. They stated that large river runoff, with a flow ratio on the order of 1.0, gives the large volume of water needed for the salt-wedge type A estuary. Smaller runoff and a flow ratio between 0.2 and 0.5 result in the partially-mixed type B estuary. The well-mixed type D estuary has a small river flow and a flow ratio less than 0.1. From the range of values of flow ratio given above for the Columbia River Estuary, it can be considered either a salt-wedge or a partially-mixed estuary.

Burt and McAlister also presented criteria for the classification of Oregon estuaries on the basis of the difference in salinity between the surface and the bottom. An estuary with a salinity difference of 20 ‰ or more is classified as type A. Estuaries with a range from 4 ‰ to 19 ‰ are of type B, and those with 3 ‰ or less are of type D. Using this method, they showed that the Columbia may be classified as type A, B, or D depending upon the discharge rate which varies throughout the year.

The classification of the Columbia River Estuary appears to be rather arbitrary as it depends both upon time and the classification method.

2. SALINITY AND CURRENT PATTERNS

As stated by Hansen (1965), the salinity intrusion is so short in length and so variable in time that a sampling pattern very dense in both time and space would be required to completely describe the longitudinal and lateral salinity distribution and their variations in this estuary. Such a dense set of samples is not available due to the difficulties of sampling in the area brought about by swift river currents and exposure to the ocean in the lower extremities of the estuary. The data which were used in the study were taken by the U. S. Army Corps of Engineers (1960) at a number of station locations as shown in Figure 1. These data, therefore, are adequate for a better description of the salinity and current distributions than is possible in most estuaries since the observations are not limited to a mid-channel line of stations. A further discussion of the data and the methods of analysis is given later.

Typical longitudinal salinity distributions for the northern channel during the period of high discharge are shown in Figures 2 and 3. As can readily be seen, the distance of the intrusion is controlled primarily by the tidal stage (during fixed runoff conditions). The fact that the estuary is quite short compared to its width is apparent when it is considered that the Columbia River varies from about 3 km to 15 km in width throughout the region of salinity intrusion while the length of the intrusion at a maximum is about 30 km. Available data indicate that, although measurable salinity intrudes as far as the Cliff Point section during the high discharge period, it rarely extends past Harrington Point for any river stage.

Some idea of the great time and space variability of salinity can be seen in Figure 4 which depicts the Clatsop Spit section. During the high discharge period, the salinity at this section varies with the tide from zero to a value approaching that of oceanic salinity. At the same time, the water along the right (or north) side is on the average somewhat less saline than that along the left (or south side). This is in accordance with the idea that the seaward flow of river water should be deflected to the right by the Coriolis effect due to the rotation of the earth. Results of calculations of lateral variations in the flux of water are tabulated later, but these results show that the overall seaward flux of water is greater on the left hand side of the channel. This indication of an inward flow on the right side (seen also in Fig. 5) is contrary to the expected influence of the Coriolis effect in estuaries and is attributed by Hansen (1965) to accelerations related to the shape of the channel. A discussion of the effects of curvature of flow on the dynamic balance of a current system in an estuary is given by Stewart (1957). Neal (1965) attributes the greater flood flow in the northern channel to the more or less direct path that it takes from the mouth. The high salinity water advected landward at depth in this region is probably due to the greater depth of the northern channel. The longer and less direct southern channel which has been dredged out to connect with the main river channel farther upstream carries the greater ebb flow.

The salinity distribution at the Clatsop Spit section and its time variations can more clearly be shown if it is plotted at equal intervals over the period of a tidal cycle. Such a representation (Fig. 6) shows that, although the surface layers are fresher on the northern side, there are greater intrusions of salt at depth on this side. On the Cliff Point

section (Fig. 1) this intrusion accounts for the presence of salt at stations D and E while salt is not found at other stations on this section. The salinity intrusion found at these two stations is below the depth of the shallow connections between the deep northern channel and the main river channel that wind between the bars and shoals found at this section. The bathymetry of the river thus prevents spillover of the intrusion into the southern channel and results in the maximum intrusion being found in the northern channel. Saline water is observed at the Cliff Point section only during times of higher high water and for a period of only about 7 hours duration. A similar representation of the velocity distribution at the Clatsop Spit section is shown in Figure 7.

The salinity distribution at the section located between Cape Disappointment and the south jetty is shown in Figure 8. It behaves in a fashion similar to that at the Clatsop Spit section. Reversal of current from flood to ebb conditions begins first on the northern side of this section (Fig. 9); the location of the one of minimum salinity coincides with that of the instantaneous ebb current maximum. The tendency for current reversal from flood to ebb at this section is opposite to that at the Clatsop Spit section (Fig. 7) where this reversal begins first on the southern side of the section. The channel curvature and local bathymetry act to control the position of the current maximum and salinity minimum.

From Figure 1, it can be seen that the Columbia River bends towards the north between the Cliff Point and Clatsop Spit sections and then reverses its curvature between the Clatsop Spit and Cape Disappointment sections. A reasonable upper limit for the value of the radius of curvature in both cases is 13.5 km. This value of radius is substituted in an expression equating the centrifugal accelerations related to the curvature

of the channel to those due to the Coriolis effect. When the equation is solved, this value of radius determines the minimum value of velocity necessary to have the accelerations related to the channel curvature greater than those due to the Coriolis effect. Instantaneous values of velocity exceeding this calculated value can then safely be concluded to result in a dominance of the curvature effects over the Coriolis effects.

Considering the Clatsop Spit section first, a velocity greater than 1.4 m/s is required on both the ebb and the flood for the accelerations related to channel curvature to be dominant. That such instantaneous velocities do exist during both ebb and flood conditions is shown in Figure 7. Using as an example a value of ebb current near the maximum for the section of 3.0 m/s, the Coriolis acceleration is $3.2 \times 10^{-4} m/s^2$ acting to the right (or north) while the centrifugal acceleration, acting towards the south since the channel is turning northward prior to reaching this section, is $6.7 \times 10^{-4} m/s^2$. The centrifugal accelerations concentrate the ebb flow on the southern side due to their dominance, but the flood is more concentrated on the northern side of the channel as the curvature is in the opposite direction when approaching this section from the seaward side. Curvature of the channel, then, accounts for the sides of the channel on which the reversals from one condition to the other (flood to ebb or ebb to flood) first take place at this section.

The Cape Disappointment section is much more complicated. At the northernmost station in this section, for example, the net flux of both water and salt is in a landward direction. This station is the only one for which this phenomenon was observed but raises the question as to whether a fixed, predominantly clockwise eddy exists in the northern region of this section. The data are insufficient to support a definite conclusion due to a lack

of good current direction data at the stations in question. Again based upon a radius of curvature of 13.5 km, a flood current magnitude of 1.4 m/s is required for the accelerations related to the channel curvature acting towards the north to overcome the Coriolis acceleration acting towards the south. The flood velocity is never this large at this section, so presumably accelerations related to channel curvature never overcome Coriolis accelerations at this section. During ebb flow they act in the same direction as Coriolis accelerations. A more reasonable radius of curvature for the northernmost station on this section, however, is 3.0 km. The flood velocity at this station does exceed the estimated value of 0.3 m/s required for the dominance of centrifugal accelerations, so channel curvature can concentrate a landward flow in the northern channel at the Cape Disappointment section.

There is a distinct tendency for the reversal of current from flood to ebb to begin first at the surface while the change from ebb to flood starts near the bottom. This conforms with the pattern for typical estuarine circulation since the general effect of salinity alone is to produce an inward velocity at the bottom and a seaward velocity at the surface. Reversal of the net flow as shown in Figure 5(c) is not observed at the Clatsop Spit section, but the low net velocity along the right-hand side of the channel appears to be associated with the salinity-caused density gradient. The net flow reversal observed at the Cape Disappointment section to the north of station C is associated with both the density gradient and the effect of channel shape on tidal motions.

Maximum ebb current observed at each of the two sections was in excess of six knots at the surface. Such large seaward velocities as are found would seem to indicate that the residence time of dissolved and suspended

materials is quite short. That such is indeed the case was concluded by Neal (1965).

Hansen (1965) gave a brief discussion of salinity and current patterns in the estuary for a period of low discharge. He found for the low-runoff season that less saline water occurs along the left bank of the estuary at the Clatsop Spit section. The net upstream flow in the lower layers characteristic of estuarine circulation also is found at this section during the period of low discharge. Hansen stated that the apparent seasonal change noted in distribution of tidal currents over the section may be as much due to a change in type of tide at time of observation as to change in river discharge. The observations for the low discharge period studied by Hansen were made during equatorial tides while the observations for the high discharge period were made during tropical tides. The salinity intrusion as was shown earlier is also controlled by the tide.

O'Brien (1952) analyzed some measurements made in 1932 by the U. S. Army Corps of Engineers in the Columbia River Estuary between Clatsop Spit and Sand Island. He also examined a period of high discharge and, in an approximate fashion, calculated the magnitude of the density current over the period of several tidal cycles as separated from the effects of tide and fresh water flow. His values are smaller than those of the fresh water flow. In effect, O'Brien showed that reversal of net current does not occur at this section during high river discharge. He did conclude, however, that the density currents can instantaneously be of the same magnitude as the tidal currents and hence play an important role in the silting and scour of the estuary. The region of stagnation where the river flow and density current are equal but in opposite directions is

associated with sediment deposits in estuaries. The large movements of this point with changing discharges and tidal conditions have caused the Corps of Engineers many problems in maintaining the present navigational depth of the Columbia River because the stagnation point also moves in response to dredging.

3. FORMULATION OF THE EQUATIONS

3.1 Fluxes Across a Vertical Cross Section

The equations to be used in conducting the analysis of the processes controlling the longitudinal flux of materials in the estuary were derived by Hansen (1965). These equations are an expansion of the approach used by Bowden (1963) in his analysis of the longitudinal salt balance. Hansen applied the equations to an analysis of the Columbia River Estuary during a period of low discharge.

Consider a left-handed coordinate system with the origin at the surface in the fresh water at the head of the estuary. The X_1 coordinate is directed down the estuary along the longitudinal axis. The lateral coordinate is X_2, and is directed horizontally from the center of the estuary toward the left-hand shore of an observer facing downstream. The X_3 coordinate points vertically downwards.

Denote by $u(x,y,z,t)$ the instantaneous seaward component of velocity at any point on a cross section normal to the longitudinal axis of the estuary. The mean value of velocity or, in the general case, any variable over the cross section is given by

$$\overline{u}(x,t) = \frac{1}{A} \iint u(x,y,z,t)\,dA \quad , \tag{1}$$

where A denotes the cross-sectional area at time t. Then, if $u_d(x,y,z,t)$ denotes the local deviation from the sectional mean,

$$u(x,y,z,t) = \overline{u}(x,t) + u_d(x,y,z,t) \tag{2}$$

and the flux of water through the section is

$$F_w(x,t) = \iint u(x,y,z,t)dA = A\bar{u}(x,t) \ . \tag{3}$$

Considering that the quantities are all changing with time, the variations can be defined in terms of three time scales which can be recognized from the data. These time scales are: (1) long period variations which appear steady during one tidal cycle, (2) large harmonic variations of tidal period, and (3) short period fluctuations which can be regarded as turbulent. The variations can be expressed in the form

$$\bar{u} = \langle\bar{u}\rangle + \bar{u}_p + \bar{u}' \ , \text{ etc. } , \tag{4}$$

where the angle brackets denote mean values over the tidal period T, i.e.,

$$\langle\bar{u}\rangle = \frac{1}{T}\int_0^T \bar{u}dt \ , \text{ etc. } , \tag{5}$$

subscript p denotes the major harmonic variations, and primes denote turbulent fluctuations.

The mean flux of water through the cross section during the tidal cycle is

$$\langle F_w\rangle = \langle A\bar{u}\rangle \ , \tag{6}$$

which, since both the velocity and cross-sectional area vary with time, becomes upon substitution from (4)

$$\langle F_w\rangle = \langle(\langle A\rangle + A_p + A')(\langle\bar{u}\rangle + \bar{u}_p + \bar{u}')\rangle \ . \tag{7}$$

Carrying out the multiplication and making the usual assumption that turbulent fluctuations are uncorrelated with the tidal period variations (Pritchard, 1954) results in

$$\langle F_w \rangle = \langle A \rangle \langle \bar{u} \rangle + \langle A_p \bar{u}_p \rangle + \langle A' \bar{u}' \rangle \; . \tag{8}$$

The term $\langle A \rangle \langle \bar{u} \rangle$ is the seaward flux of water through the mean cross-sectional area carried by the mean flow. This quantity is usually referred to as the nontidal drift. The flux $\langle A_p \bar{u}_p \rangle$ results from the tide and will generally be landward. Figure 10(a), (b) shows that the tidal current-height relation in the Columbia River Estuary is that of a wave nearly midway between a pure standing wave and a pure progressive wave. This same relationship is also true during conditions of low discharge as indicated by both Hansen (1965) and Neal (1965). The turbulent flux, $\langle A'u' \rangle$, representing water transport by motions of less than semidiurnal period, is negligible (Hansen, 1965).

Neglecting molecular diffusion, the instantaneous flux of salt through a vertical cross section perpendicular to the mean flow is given by

$$F_s(t) = \iint usdA = A(\bar{u}\,\bar{s} + \overline{u_d s_d}) \quad , \tag{9}$$

with the quantities denoted as before.

The term $A\bar{u}\,\bar{s}$ expresses the instantaneous salt flux due to the sectional mean current and salinity. This term, therefore, reverses direction in the presence of a tidal current, but, if conditions were strictly stationary, it would represent the seaward salt advection by the freshwater discharge mode of circulation. The term $A\overline{u_d s_d}$ represents a modification to the above by a correlation between the local deviations of salinity and current from their respective means. Again considering strictly stationary conditions, the second term would represent an upstream salt flux due to gravitational convection currents. This second term has been called the "shear effect" and arises in estuaries primarily from density currents and from velocity

shear induced in tidal currents by bottom friction. At low river discharge, Hansen (1965) found the extremes of this term coincided with the extremes of tidal current and, therefore, with the extremes of tidal current shear. He suggested that there is a definite periodic variation due to velocity shear superimposed upon a nearly constant effect of density current. The separation is also apparent at the time of high discharge as may be seen from Figure 10(d).

Since the conditions are not expected to be stationary in this estuary for periods less than a complete tidal cycle of period T, the mean salt flux over a tidal cycle is

$$\langle F_s \rangle = \frac{1}{T} \int_0^T F_s(t)dt = \langle A\bar{u}\bar{s} \rangle + \langle A\overline{u_d s_d} \rangle . \tag{10}$$

Considering that the velocity, salinity, mean deviation product of velocity and salinity, and cross-sectional area are all variable in time, the average salt flux contains terms derived from

$$\begin{aligned} \langle F_s \rangle = {} & \langle (\langle A \rangle + A_p + A')(\langle \bar{u} \rangle + \bar{u}_p + \bar{u}')(\langle \bar{s} \rangle + \bar{s}_p + \bar{s}') \\ & + \langle (\langle A \rangle + A_p + A')(\langle \overline{u_d s_d} \rangle + (\overline{u_d s_d})_p + (\overline{u_d s_d})') \rangle . \end{aligned} \tag{11}$$

Upon expanding (11) in the same fashion as was done to obtain equation (8) from (7), again making the assumption that the turbulent fluctuations are uncorrelated with tidal period variations, and substituting equation (8), equation (11) is finally reduced to the form

$$\begin{aligned} \langle F_s \rangle = {} & \langle F_w \rangle \langle \bar{s} \rangle + \langle A \rangle \langle \bar{u}_p \bar{s}_p \rangle + \langle \bar{u} \rangle \langle A_p \bar{s}_p \rangle \\ & + \langle A_p \bar{u}_p \bar{s}_p \rangle + \langle A \rangle \langle \overline{u_d s_d} \rangle + \langle A_p (\overline{u_d s_d})_p \rangle \\ & + \langle \langle A \rangle (\bar{u}'\bar{s}') + \langle \bar{u} \rangle (A'\bar{s}') + A'(\bar{u}'\bar{s}' + (\overline{u_d s_d})') \rangle . \end{aligned} \tag{12}$$

These terms, in order, are the seaward flux associated with:

(1) the mean river flow and mean salinity;

(2) correlation of tidal period variations of the sectional mean salinity and current;

(3) correlation between tidal period variations of cross-sectional area and sectional mean salinity;

(4) third order correlation of tidal period variations in mean sectional salinity, mean sectional velocity, and cross-sectional area;

(5) the mean shear effect;

(6) covariance of shear effect and cross-sectional area;

(7) correlation between shorter period fluctuations of the various quantities which are:

a.) short period variations of the sectional mean salinity and current;

b.) short period variations of cross-sectional area and sectional mean salinity;

c.) third order correlation of short period variations in salinity, current, and cross-sectional area;

d.) covariance of short period variations in cross section and shear effect.

Equation (12) does not completely separate the advective processes from the tidal processes. The reason for this is that the shear effect term $A\overline{\overline{u_d s_d}}$ shown in equation (9) is due both to density currents and to velocity shear induced in the tidal currents by bottom friction. As discussed earlier, the term $A\bar{u}\bar{s}$ in equation (9), under strictly stationary conditions, would represent the seaward salt advection by the freshwater discharge mode of circulation while the shear effect term $A\overline{\overline{u_d s_d}}$ would

represent an upstream salt flux due to gravitational convection currents under the same conditions. Assuming that such conditions do exist for averages over a complete tidal cycle, a method of estimating the magnitude of the density currents is available.

The total net advective salt flux can be calculated by integrating the product of the local mean current, the local mean salinity, and the incremental area through which this local flux is taking place over the cross-sectional area. The seaward advective salt flux due to the mean velocity is equal to the product $\langle A \rangle \langle \bar{u} \rangle \langle \bar{s} \rangle$ where, as before, the bar indicates a sectional average. The upstream salt flux due to the density current is obtained as the difference between the net advective salt flux and the salt flux due to the mean flow.

With the values of total net salt flux (which is equal to zero for steady state conditions) and turbulent salt flux computed from equation (12) together with the value of net advective salt flux just calculated, the net salt flux due to tidal processes can be estimated as it is the difference required to make up the balance. The time average $\langle A\bar{u}_p \bar{s}_p \rangle$ gives the salt flux due to the sectional average tidal current. The difference between this value and the value of net salt flux due to tidal processes gives the magnitude of the effects of other processes, including friction, on the tidal variations.

3.2 Fluxes Across a Vertical Section of Unit Width

The use of average values taken over an entire cross section to compute the fluxes across the section tends to obscure variations which might take place either laterally or vertically. Some idea of the lateral variations can be obtained by combining the approaches of Bowden (1963), who took a

longitudinal salt balance across a section of unit width perpendicular to the mean flow at each station examined in the River Mersey, and Hansen (1965). Since the U. S. Army Corps of Engineers (1960) data were taken for the Columbia River using a number of stations at each cross section, lateral variations can be shown by computing fluxes at each individual station and comparing the values obtained with those for the cross section as a whole.

The development of the expression for flux of salt across a section of unit width parallels that of the expression for fluxes across the entire cross section. The depth of water, h, times the unit width assumed for dimensional considerations gives the cross-sectional area across which the flux is computed. Variations of water depth of periods less than tidal are considered to be negligible. The final expression is

$$\langle Q_s \rangle = (\langle h \rangle \langle \bar{u} \rangle + \langle h_p \bar{u}_p \rangle) \langle \bar{s} \rangle + \langle h \rangle \langle \bar{u}_p \bar{s}_p \rangle$$
$$+ \langle \bar{u} \rangle \langle h_p \bar{s}_p \rangle + \langle h_p \bar{u}_p \bar{s}_p \rangle + \langle h \rangle \langle \overline{u_d s_d} \rangle$$
$$+ \langle h_p (\overline{u_d s_d})_p \rangle + \langle h \rangle \langle \bar{u}' \bar{s}' \rangle . \qquad (13)$$

The terms correspond with those given for fluxes across the total cross section. The space averages are taken over the vertical at the individual station in this instance rather than over the whole cross section.

This approach separates the salt flux mechanism better than does the approach of Bowden (1963). In the notation of this paper, with R taken to be the river discharge, Bowden's equation is composed of the following terms:

$Q_1 = \langle h \rangle \frac{R}{A} \langle \bar{s} \rangle$ which is the flux due to advection by the mean flow corresponding to the river discharge;

$Q_2 = \langle h \rangle \langle \bar{u}_p \bar{s}_p \rangle + \langle h_p \bar{u}_p \bar{s}_p \rangle$ which are some of the correlations in time of variations of cross section and depth-mean salinity and velocity;

$Q_3 = \langle h \rangle \langle \overline{u_d s_d} \rangle + \langle h_p (\overline{u_d s_d})_p \rangle$ which is the shear effect and the covariance of depth and depth-mean shear effect;

$Q_4 = \langle h \rangle \langle \overline{u}'\overline{s}' \rangle$ which does not correspond exactly with any of the terms but is the same general idea of a transport of salt by turbulent fluctuations.

Upon comparison it can be seen that Bowden did not make a good separation of the river discharge term. It was his definition of the mean flow corresponding to the river discharge as $\frac{R}{A}$, however, which allowed him to give a correct expression for the net flux due to this flow. Pritchard (1958) discussed the uses and misuses of such a definition. Bowden also did not include the $\langle \bar{u} \rangle \langle h_p \bar{s}_p \rangle$ term which is one of the time correlations of variation of depth and depth-mean salinity.

3.3 Effective Coefficient of Horizontal Eddy Diffusion

Taking sectional average values over a tidal period for the salinity and velocity has reduced the problem to one dimension. This approach allows an estimation of the effective coefficient of horizontal eddy diffusion when two assumptions are made. The transport of salt required to balance the net advective salt flux is regarded as due to the longitudinal eddy diffusion, and conditions averaged over a tidal period are assumed not to be changing with time. The details of the mixing processes across the vertical section are neglected.

The total net advective salt flux, representing the sum of the seaward salt flux due to the freshwater discharge mode and the upstream salt flux

due to the density current, is calculated by integrating the product of the local mean current $\langle u \rangle$, the local mean salinity $\langle s \rangle$, and the incremental area dA through which this local flux is taking place. Applying the principle of conservation of salt, which states that the steady-state net salt flux across any cross section of the estuary is zero, results in

$$K_x = \frac{\iint \langle u \rangle \langle s \rangle \, dA}{\langle A \rangle \dfrac{d\langle \bar{s} \rangle}{dx}} \quad . \tag{14}$$

The term K_x is the effective coefficient of horizontal eddy diffusion with the remaining terms the same as in earlier usage.

The coefficient K_x may be calculated for any position in the estuary as long as the cross-sectional area and the distributions of salinity and current are known. As stated by Pritchard (1958), the coefficient obtained in these calculations does not represent the mean value of the diffusion coefficient. It merely is a function which, when substituted into equation (14), can be used to describe the distribution in time and space of the mean salt content. Since similar coefficients have been estimated for other estuaries by Kent (1958), Hansen and Rattray (1965), Bowden (1963), and others, the coefficient can be used as a basis for comparison with these other estuaries.

3.4 Effective Coefficient of Vertical Eddy Diffusion

Pritchard (1954) gives an equation for the continuity of salt within an element of volume extending across an estuary of variable width, but his equation assumes lateral homogeneity which is not found in the Columbia. For such an element where lateral variations are considered, it can be shown that

$$\frac{\partial s}{\partial t} = -\frac{1}{W}\frac{\partial}{\partial x}(W\,\overline{us}_y) - \frac{1}{W}\frac{\partial}{\partial z}(W\,\overline{ws}_y) \quad , \tag{15}$$

where W is the width of the estuary at depth Z. The bar indicates a space average while the subscript y denotes that the average is taken over the lateral coordinate. The term $(\overline{us}_y)$, however, will not account for the turbulent flux in the Columbia as the measurement procedure filters the turbulent components out of the u and s determinations. These turbulent components were found to be relatively small for the cross section as a whole and therefore will be neglected for each depth interval.

The average value of any variable within the element is given by an expression of the form

$$\overline{u}_y(x,z,t) = \frac{1}{W}\int u(x,y,z,t)\,dW \quad , \tag{16}$$

The time variations of $\overline{u}_y$ can be expressed in a form similar to equation (4)

$$\overline{u}_y = \langle\overline{u}_y\rangle + \delta\overline{u}_y \quad , \tag{17}$$

where $\delta\overline{u}_y$ includes all time dependent fluctuations.

Substituting expressions of the form of (17) into equation (15) and taking the time mean gives

$$\left\langle\frac{\partial s}{\partial t}\right\rangle = -\frac{1}{W}\frac{\partial}{\partial x}(W\langle\overline{us}_y\rangle) - \frac{1}{W}\frac{\partial}{\partial z}(W\langle\overline{ws}_y\rangle) \quad , \tag{18}$$

where the brackets signify the time averages.

Equation (18) is the expression for the salt balance in an element at a particular depth. To find the salt balance across the entire cross section, this equation must be integrated over the vertical. Carrying out this integration gives

$$- \int_0^z \frac{1}{W} \frac{\partial}{\partial x} (W \langle \overline{us}_y \rangle) dz - \langle \overline{ws}_y \rangle = 0 \; , \tag{19}$$

where it is assumed that the transfer of salt across the entire cross section is in a steady state condition.

The term $\langle \overline{ws}_y \rangle$ is the total vertical salt flux which includes the vertical advection of salt. The turbulent vertical salt flux is the difference between $\langle \overline{ws}_y \rangle$ and the vertical advective salt flux. Upon substituting for the vertical eddy flux term the product of a diffusion coefficient and a spatial derivative of the mean salt concentration, the final expression for the effective coefficient of vertical eddy diffusion becomes

$$K_z(z) = \frac{\langle \overline{w}_y \rangle \langle \overline{s}_y \rangle + \int_0^z \frac{1}{W} \frac{\partial}{\partial x} (W \langle \overline{us}_y \rangle) dz}{\frac{d \langle \overline{s}_y \rangle}{dz}} \; . \tag{20}$$

4. DATA SOURCES AND ANALYSIS

4.1 Sources

Data used in the study were taken by the U. S. Army Corps of Engineers in 1959 and published in an interim report in 1960. The data consist primarily of continuous graphs of both current speed and salinity over the period of one tidal cycle. Direction data for the current are also given, but they are not so complete as those for the current speed and salinity because a great number of readings were missed for various reasons. No temperature data were published. Tidal curves for the period of observations are also available which allows the cross-sectional area of the river to be estimated.

The measurement program was carried out in three nine-day periods covering conditions of low, intermediate, and high river discharge. Measurements were made at seven cross sections extending upstream approximately 100 km, but only the lower three cross sections were in the region of salinity intrusion. This study is concerned only with the period of high discharge for which measurements were taken June 16 through June 25 when the river flow ranged between 15000 m^3/s and 16300 m^3/s.

Measurements were made three feet below the water surface and at the quarter, half, three-quarter, and two-feet-above-the-bottom depths at all stations. Data at each station were taken only over the period of one tidal cycle. This restricts the data to some degree as the length of the record is approximately equal to the period of one of the major fluctuations

observed, namely the diurnal tidal component. The information was gathered at one-half hour intervals placing a lower limit on the time scale which can be considered even though continuous curves were plotted versus time.

The three cross sections which were metered in the region of salinity intrusion and the station locations for each are shown in Figure 1. Further details of the measurement program are available in the interim report of the U. S. Army Corps of Engineers (1960, Vol. I).

4.2 General Reduction of Data

Values of salinity, current speed, and tidal height were read from the data curves at intervals of one lunar hour over the period of one tidal cycle. Using the tidal height information in conjunction with plots of the cross section of the river that were obtained from U. S. Coast and Geodetic Survey Chart No. 6151, the cross-sectional area was planimetered at each section, for each time of observation. Tidal time differences between the measurement of a tidal height at the tide recording gauges and the actual occurrance of this height at a section were determined to be small and were neglected. Appreciable errors in the computed fluxes can be introduced only if the time lag is comparable with the interval between observations or if there is a large change in cross-sectional area during this period. In these cases, evaluation of terms involving tidal variations of cross-sectional area must take into account this time delay.

Values of salinity and current speed were observed at five different depths at each station. These values were averaged numerically both over the vertical at each station for each time of observation as well as at the individual depth at each station over the tidal cycle. An average direction correction, determined from the current direction data, was applied to the

current speed so that the component of velocity normal to the cross section could be obtained at each station. It is the normal component of velocity that must be used in the computation of salt and water fluxes across the vertical section. The use of an average direction correction appeared to be a reasonable approach as the velocity vectors during both the flood and ebb only fluctuated within reasonably narrow limits once the condition had been established. In general, the ebb and the flood velocities had to be corrected separately as their dominant directions usually deviated from the normal to the cross section by differing amounts.

The product of the deviation of salinity from the mean with the corresponding deviation of velocity was calculated at each depth at each time and then averaged over both the vertical and then the tidal cycle at each station to obtain the shear effect term. The shear effect is defined as a modification to the instantaneous salt flux resulting from a correlation between the local deviations of current and salinity from their respective sectional mean values. It arises primarily from density currents and from velocity shear induced in tidal currents by bottom friction.

Cross-sectional averages of salinity, velocity, and shear effect were calculated by numerically averaging the values of these quantities at the stations shown on the individual cross sections (Fig. 1). The values were weighted in proportion to the amount of the cross section represented by each station to avoid over weighting the data to one particular region of the channel. In other words, the values at a station representative of two-thirds of the cross-sectional area would be doubled before being summed with a station representative of one-third of the total area. The sum in this example would be divided by three to obtain the sectional average value of the quantity being considered.

Tha average quantities calculated on a sectional basis and on an individual station basis were subjected to multiple regression analysis by computer to determine harmonic expressions of the form

$$P = P_0 + P_1 \cos(\omega_1 t - \phi_1) + P_2 \cos(\omega_2 t - \phi_2) \qquad (21)$$

which would be representative of the variations of these quantities. A plot of the smooth curves obtained superimposed on the observed data points is shown in Figure 10 for the Clatsop Spit section as a typical representative.

Since it was felt that the fluctuations were largely a result of the tide, the harmonic expressions use the frequencies of the K_1 and M_2 harmonic tidal constituents. The expressions obtained were found to adequately describe the cross-sectional area and tidal height terms, the average velocity, and the average salinity. Expressions using the frequencies of the four major harmonic tidal constituents (K_1, M_2, O_1, and S_2) were found to make no significant improvement in the degree of approximation probably due to the closeness of the two additional frequencies to those which were already present and the shortness of the record. The addition of two frequencies merely adds two more cosine terms to the expression shown in equation (21). This addition of terms complicates the averaging of the products of the quantities represented with no significant gain in accuracy, so the two term approximation is the one which is used.

The shear effect terms are not so well described by the expressions using only the frequencies of the K_1 and M_2 harmonic tidal constituents as shown in Figure 10(d). Since a multiple regression analysis computer program had already been developed to handle four frequencies as discussed above, two frequencies (other than O_1 and S_2) were selected for use in addition to the frequencies of the K_1 and M_2 harmonic tidal constituents.

The two frequencies chosen were that of twice the frequency of the M_2 constituent and that of three times the frequency of the K_1 component. Selection of these frequencies was based upon both a mathematical consideration of the frequencies that might be expected to result from the product of two two-component quantities and the assumption that the local deviations of salinity and velocity, which were the two quantities under consideration, exhibited the same periodicity shown by their respective sectional mean values. These additional frequencies look very similar to those of constituents due to the distortional effects of shallow water on tidal motion.

The resultant expression using the four harmonic terms is a good approximation to the observed variations of the shear effect. The values of P_0, P_1, and P_2 in equation (21), which are coefficients that are duplicated between the two-frequency and four-frequency expressions, are nearly the same in both expressions. The two additional coefficients in the four-frequency expressions, P_3 and P_4, contribute negligibly to the calculated fluxes when all quantities other than shear effect are expressed using two-frequency approximations. Thus, the four-frequency expressions are considered as merely better approximations to the observed shear effect data as they do not significantly modify the fluxes calculated using the two-frequency approximation for shear effect. Since these four-frequency expressions were not used in the final computations, they will not be shown.

The data for each individual station on each cross section were analyzed in a similar fashion. Cross-sectional area terms were replaced with tidal height terms since sections of unit width perpendicular to the flow were assumed at each station to provide the area across which the fluxes at each

station could be calculated directly. Examination of the stations individually allowed some idea of lateral variations in the estuary to be obtained.

The Cliff Point section was the most difficult to treat. During the high discharge period, saline water is found at this section only at the higher high water stage and then only for about seven hours duration. Thus, this section was for the most part neglected, partly as a consequence of its location and partly due to the complexity of the cross section found there.

5. DISCUSSION OF RESULTS

5.1 Fluxes Across a Vertical Cross Section

The data were submitted to multiple regression analysis by computer to determine harmonic expressions of tidal frequency suitable for substitution into equations (8) and (12). Figure 10 shows representative curves which have the following equations:

$$\left.\begin{aligned}
&\text{(a)}\ \langle A \rangle + A_p = [3.85 + 0.24\ cos(0.2717t - 0.6140) \\
&\qquad\qquad + 0.47\ cos(0.5236t - 2.8933)]10^4 m^2 \\
&\text{(b)}\ \langle \overline{u} \rangle + \overline{u}_p = [0.39 + 0.40\ cos(0.2717t - 2.5163) \\
&\qquad\qquad + 1.26\ cos(0.5236t + 1.2293)]m/s \\
&\text{(c)}\ \langle \overline{s} \rangle + \overline{s}_p = [10.22 + 4.10\ cos(0.2717t - 0.3967) \\
&\qquad\qquad + 10.75\ cos(0.5236t - 3.1416)]kg/m^3 \\
&\text{(d)}\ \langle \overline{u_d s_d} \rangle + (\overline{u_d s_d})_p = [-\ 0.96 + 0.0\ cos(0.2717t - 0.0) \\
&\qquad\qquad + 2.28\ cos(0.5236t - 1.5708)]kg/m^2s\ .
\end{aligned}\right\} \quad (22)$$

Values for mean seaward water flux are obtained by substituting equations of the form of (22a) and (22b) into (8) and averaging over the tidal cycle as shown in (5). The results of the calculation of mean seaward water flux are shown in Table I.

The river hydrograph (U. S. Army Corps of Engineers, 1960, Vol. IV) indicates a median discharge of about 15600 m^3/s during the period of

observations. Accuracy of the hydrograph is considered no better than fair to good since the Columbia is not gauged below the Dalles and only 50 percent of the drainage basin runoff into this part of the river is gauged on a regular basis (Budinger, Coachman, and Barnes, 1964).

Table I

Mean seaward water flux across vertical cross section (10^4 m^3/s)

	$\langle A \rangle \langle \bar{u} \rangle$	$\langle A_p \bar{u}_p \rangle$	$\langle F_w \rangle$
Cape Disappointment section	1.93	-0.15	1.78
Clatsop Spit section	1.50	-0.16	1.34

The computed net transport at the Cape Disappointment section exceeds the median river hydrograph value for that time period by about 11%. The U. S. Army Corps of Engineers (1960, Vol. I) stated that the current meters which were used had a known tendency to over-register in flow conditions containing appreciable vertical components. Since the Cape Disappointment section is located in waters known to be rough, the boats used to make the observations were wave-tossed. A noticeable increase in the frequency of meter clicks occurring in sequence with the rising and falling of the boat as swells passed was found. At the Clatsop Spit section, however, the computed net transport is less than the median river hydrograph for the time of observations by about 14%. A possible cause for this is that the average direction correction may be slightly in error. Also, an unknown but probably small amount of water flows behind the sand island at the northern end of the section. Overall agreement between the computed values and the river hydrograph is considered good.

During the high discharge condition, about 90% of the metered nontidal drift is due to the freshwater discharge as compared with about 70% observed by Hansen (1965) for the low runoff condition. The remainder in both cases is a compensation current for the inward transport by the tidal currents.

The contributions to the salt flux as given in equation (12) are evaluated in a similar fashion and are shown in Table II.

Table II

Mean seaward salt flux across vertical cross section (10^5 kg/s)

Cape Disappointment section:	
$\langle F_w \rangle \langle \bar{s} \rangle$	2.58
$\langle A \rangle \langle \overline{u_p s_p} \rangle$	-0.54
$\langle \bar{u} \rangle \langle A_p \bar{s}_p \rangle$	0.11
$\langle A_p \bar{u}_p \bar{s}_p \rangle$	-0.07
$\langle A \rangle \langle \overline{u_d s_d} \rangle$	-0.85
$\langle A_p (\overline{u_d s_d})_p \rangle$	-0.01
Clatsop Spit section:	
$\langle F_w \rangle \langle \bar{s} \rangle$	1.36
$\langle A \rangle \langle \bar{u}_p \bar{s}_p \rangle$	-0.88
$\langle \bar{u} \rangle \langle A_p \bar{s}_p \rangle$	0.10
$\langle A_p \bar{u}_p \bar{s}_p \rangle$	-0.07
$\langle A \rangle \langle \overline{u_d s_d} \rangle$	-0.37
$\langle A_p (\overline{u_d s_d})_p \rangle$	0.03

The sum of the fluxes given by the table must equal the difference between the change in salt storage above the section during the period and the flux through the section due to the turbulent fluctuation terms which are shown in equation (12). Computation of either of these separately is not possible with the present data, but a qualitative estimate can be made concerning the relative sizes of these terms based primarily upon the river hydrograph.

The sum of the fluxes for the Cape Disappointment is 1.22×10^{-5} kg/s which shows a net seaward flux and represents 45% of the mean seaward salt flux through the section. To balance this, the sum of the fluxes due to the turbulent fluctuations and the change in salt storage above the section must equal -1.22×10^{-5} kg/s. The fact that a steady state situation did not exist in the river during the taking of the data is indicated by the river hydrograph. The river hydrograph shows that the river discharge for the period one day prior to and during the period of observations at this section increased from 15100 m^3/s to 16300 m^3/s. Estimates of the flushing time made by Neal (1965), although made for discharges of less than 10700 m^3/s, give values as low as two tidal cycles for the flushing time of the estuary. This indicates that the residence time of dissolved or suspended materials in the estuary must be relatively short. Consequently, the major part of the term which is the summation of the fluxes due to turbulent fluctuations and the change in salt storage is probably due to a decrease in salt storage above the section. It is felt that such is the case because of the fairly large change in the river flow.

Since the river discharge decreases during the two day period prior to the taking of observations at the Clatsop Spit section, the value of the

term required to make a balance at this section, $-.17 \times 10^5$ kg/s, is interpreted as being due to a landward flux by turbulent fluctuations. In this case the term only represents 11% of the mean seaward salt flux through the section. At both sections, then, the turbulent transport term is considered to be of secondary importance in the salt balance.

Assuming stationary conditions over the period of a tidal cycle, the salt flux due to tidal processes can be completely separated from that due to the advective processes corresponding to the freshwater discharge and gravitational convection current modes as discussed during the formulation of the equations. The results of this separation are shown in Table III.

Despite the stratification of the Columbia River, the influence of density currents in maintaining the salt balance appears to be small relative to tidal processes. In fact, the mean current at the Clatsop Spit section (Fig. 5) advects salt outward at all depths. The low net velocity along the right-hand side of this section appears to be associated with a salinity-caused density current that is dominated by the large river discharge over the majority of the tidal cycle.

At the Clatsop Spit section, the density current balances only about 17% of the salt advected seaward by the mean river discharge. The exact percentage is difficult to determine at the Cape Disappointment section as it is not known how much of the salt advected seaward through this section is not balanced by landward transport as a result of a change in the upstream storage.

Assuming that the great difference between landward and seaward salt flux at the Cape Disappointment section is almost entirely due to change in salt storage above this section and that conditions are nearly stationary

Table III

Processes contributing to salt flux across vertical cross section (10^5kg/s)

Cape Disappointment section:	
Advective	2.73
River discharge	3.18
Density current	-0.45
Tidal	-1.13
Average tidal current	-0.86
Other processes (including friction)	-0.27
Turbulent and change in storage	-1.22
Clatsop Spit section:	
Advective	1.27
River discharge	1.53
Density current	-0.26
Tidal	-1.10
Average tidal current	-1.15
Other processes (including friction)	0.05
Turbulent and change in storage	-0.17

at the Clatsop Spit section, some observations can be made concerning the magnitudes of the various processes contributing to the maintenance of the salt balance. At the Clatsop Spit section, about 60% of the salt advected seaward by all processes is balanced by covariance between fluctuations of tidal period, about 25% (of which about 3/4 is due to density currents) is balanced by shear effects, and the remaining 15% is balanced by fluctuations of tidal period and shorter. Hansen (1965), for the low discharge period

at this same section, showed that of the salt advected seaward about 40% is balanced by covariance between fluctuations of tidal periodicity, about 45% is balanced by shear effects, and the remaining 15% is balanced by short-period fluctuations. As a comparison with the above results for the Columbia, horizontal advection was found to provide over 95% of the landward salt flux in the James River estuary (Pritchard, 1954). The density current structure in this estuary is quite weak despite its stratification. The combination of the stratification and the depth of the estuary, however, allows the density current, even though poorly developed, to be of significance in the balance.

Of the salt advected seaward by all processes at the Cape Disappointment section, but excluding that estimated to be flushed from the estuary due to the increasing river flow prior to and during the time of observations, about 30% is balanced by covariance between fluctuations of tidal period, about 50% (of which about 1/2 is due to density currents) is balanced by shear effect, and the remaining 20% is balanced by fluctuations of tidal period and shorter. The relative importance of the major terms is seen to be variable from section to section during the same discharge period. This is possibly a function of the locations of the sections and their proximity to the ocean.

5.2 Fluxes Across a Vertical Section of Unit Width

Harmonic expressions similar to equation (22) were obtained from multiple regression analysis by computer for the quantities at individual stations on the cross sections. Station depths obtained by adding the tidal heights to the depth at mean lower low water were used to determine areas

at the stations across which the fluxes take place. Since unit width is assumed for the vertical section, the magnitude of the instantaneous station depth is numerically equal to the area of the section under consideration.

Values for the mean seaward water flux are obtained by substituting into equation (8) and averaging over the tidal cycle as shown in (5). The results of the calculations are shown in Table IV.

Table IV

Mean seaward water flux across vertical section of unit width (m^3/s)

		$\langle h \rangle$	$\langle h \rangle \langle \bar{u} \rangle$	$\langle h_p \bar{u}_p \rangle$	$\langle Q_w \rangle$
Cape Disappointment section:					
Station	A	7.0	3.0	-0.4	2.6
	B	16.2	10.2	0.0	10.2
	C	8.9	5.4	-0.8	4.6
	D	8.3	-4.2	0.1	-4.1
Clatsop Spit section:					
Station	A	12.2	7.6	-0.4	7.2
	C	22.0	8.3	-0.5	7.8
	D	15.3	2.6	-0.5	2.1

Large lateral variations can be seen in the flux of water. Differences in depth at each station should be considered, and average station depths, $\langle h \rangle$, are shown in meters. Quite naturally, deeper stations are expected to carry greater flow assuming a uniform flow rate, but the deeper stations here are also located in the center of the channel where the flow is the greatest (Figs. 1, 7, and 9). The greater flux in the channel center with regions of lesser seaward flux toward either side as expected is apparent from the data.

The fluxes at the Clatsop Spit section show that the southern channel carries a greater seaward flow than the northern channel. Since Station D in the northern part of this section is not in shallow water, the low value of mean flux found is a result of landward flow here for part of the tidal cycle. That such a flow does exist on the northern side of the estuary is shown by an examination of the variations across the Cape Disappointment section. At the northernmost station on the Cape Disappointment section, the net flux of water is in a landward direction which is probably due to a combination of the strength of the tidal forces and the local morphology of the estuary. This is the only station for which this phenomenon was observed, and it corroborates the finding at the Clatsop Spit section of an inward flow concentrated in the northern channel for part of the tidal cycle. At all of the stations except the northernmost one on the Cape Disappointment section, the freshwater discharge is generally 85% or more of the metered nontidal drift.

The contributions to the salt flux as given in equation (13) are evaluated in a similar fashion and are shown in Table V.

A balance to the salt transport equation is neither expected nor achieved at each station since conditions were not uniform across the channel and were not stationary over the tidal cycle as mentioned previously.

The calculation of lateral variation in the fluxes reveals a situation not normally recognized in estuaries. The Columbia River, which is divided into two channels about 10 km upstream from the mouth, exhibits a greater flood flow in the northern channel while the southern channel carries a greater ebb flow. This cross-channel flow differential is attributed primarily to accelerations related to the shape of the channel and was discussed earlier in a quantitative fashion. The variations in salt flux

conform to the variations in the flux of water.

Table V

Mean seaward salt flux across vertical section of unit width (kg/s)

Cape Disappointment section:

Station	A	B	C	D
$Q_w \quad \bar{s}$	29.7	164.0	65.2	-67.2
$h \quad \bar{u}_p \bar{s}_p$	-1.5	-49.4	-27.8	15.2
$\bar{u} \quad h_p \bar{s}_p$	1.8	3.8	4.4	-2.4
$h_p \bar{u}_p \bar{s}_p$	-0.3	-4.1	-4.2	-2.5
$h \quad \overline{u_d s_d}$	-9.7	-36.3	-21.1	-5.9
$h_p (\overline{u_d s_d})_p$	0.2	-0.5	-0.4	-0.3

Clatsop Spit section:

Station	A	C	D
$Q_w \quad \bar{s}$	79.5	88.2	18.5
$h \quad \bar{u}_p \bar{s}_p$	-23.0	-77.3	-35.2
$\bar{u} \quad h_p \bar{s}_p$	4.6	3.2	1.0
$h_p \bar{u}_p \bar{s}_p$	-2.6	-4.9	-3.2
$h \quad \overline{u_d s_d}$	-18.5	-29.0	-1.1
$h_p (\overline{u_d s_d})_p$	0.4	0.3	0.2

5.3 Effective Coefficient of Horizontal Eddy Diffusion

Since data are available at only three cross sections spaced longitudinally, good estimates of the value of salinity gradient needed to substitute into equation (14) can be obtained only at the central section of the three.

The central or Clatsop Spit section is the most logical area in any case to examine the values of the effective coefficient of horizontal eddy diffusion. The seaward section at Cape Disappointment is close to the mouth with resultant domination by the ocean while the Cliff Point section is dominated by the river and reached only by a short duration salinity intrusion extending over less than half its width. Also, conditions at the Clatsop Spit section nearly approximate a steady-state situation. The eddy coefficient calculation using equation (14) is shown in Table VI. The sections are referred to by numbers increasing consecutively upstream from the mouth.

Table VI

Calculation of K_x for Clatsop Spit section

	Section 1		Section 2		Section 3
$\iint u \quad s \quad dA (10^5 \mathrm{kg/s})$			1.27		
$A \ (10^4 \mathrm{m}^2)$			3.85		
$\overline{s} \ (\mathrm{kg/m}^3)$	14.6		10.4		0.9
$d \ \overline{s} \ (\mathrm{kg/m}^3)$		4.2		9.5	
$dx (\mathrm{km})$		5.41		14.46	
$\frac{d \ s}{dx} (10^{-3} \mathrm{kg/m}^4)$		.776		.656	
$K_x (10^3 \mathrm{m}^2/\mathrm{s})$			4.6		

The coefficient calculated in Table VI is based on the average value of the two gradients shown. The gradient obtained in this fashion is considered to be the best approximation possible from the data.

Some errors are introduced into the approximation due to the non-steady state conditions resulting from changes in the river flow. Data at the Cape Disappointment section (Section 1) were taken when the discharge was significantly higher than the median river hydrograph for the period of observations at all the sections. The mean salinity at this section is consequently somewhat lower than it would have been had the river been at steady state resulting in a smaller computed value of K_x than might be expected during steady state. The value 4600 m^2/s compares very favorably, however, with the value of 5100 m^2/s estimated by Hansen and Rattray using a different method. Data at Sections 2 and 3 were taken with the river discharge near the median river hydrograph so that conditions approximated a steady state at these sections.

The values of flux associated with the different processes contributing to the salt balance for the Columbia River as previously discussed confirm that the advection of salt by the nontidal circulation alone cannot maintain the longitudinal balance even assuming a correction for advective flux due to a change in storage of the estuary. Since the density-current structure is weak, despite a sometimes strong vertical stratification, the longitudinal salt balance must also be maintained by an upstream flux of salt associated with tidal-period fluctuations. For this reason, the value of the effective coefficient of horizontal eddy diffusion estimated in the classical fashion is extremely large when compared with similar values estimated for other estuaries by other owrkers. The coefficient for the Columbia is 10 to 100 times the values given for several British estuaries, including the River Mersey, by Bowden (1963), 10 times the values given for the Delaware River by Kent (1958), and 100 times the estimate given for the James River by Hansen and Rattray (1965).

5.4 Effective Coefficient of Vertical Eddy Diffusion

The fact that data were taken over a range in depth allowed substitution into equation (20) so that the effective coefficient of vertical eddy diffusion could be determined at several depths. Only the Cape Disappointment and Clatsop Spit sections were used in this calculation due to the complexity of the cross section and the limited duration of the salinity intrusion at the Cliff Point section which were mentioned earlier.

A plot of salt transport versus depth showed that the Clatsop Spit section was nearly in a steady state condition. A similar plot at the Cape Disappointment section showed a large net seaward flux which has been attributed to a change in salt storage above the section as a result of increasing river discharge prior to and during the period of the taking of observations. The transport at this section was balanced using two independent approaches to approximate a nearly steady state condition at this section as well. Values obtained for the vertical coefficient utilizing the forced steady state conditions are shown in Table VII. These estimated values are comparable to the range of values found in other estuaries. The values are also comparable with the range of 15 - 30 cm^2/s estimated at the time of high discharge by Hansen and Rattray (1965) using a different method.

The method of approximating a steady state condition for salt transport influences the results of the calculations. One method involved uniformly correcting the average velocity profile to reduce volume transport to the median river hydrograph value followed by subtracting the salt transport attributed to change in storage above the section. The second method adjusted the vertical distribution of salt transport to conform qualitatively

to that at the Clatsop Spit section utilizing the ratio of the area of the layer being considered to total cross-sectional area. The general agreement of the results, in magnitude if not in vertical distribution, gives some confidence to the calculation.

The low relative value for the coefficient at the 5-meter depth is felt to be due to a combination of the stratification of the Columbia River and the nature of the cross section.

Table VII

Estimate of K_z for the Lower Columbia

z(m)	K_z (cm^2/s)	
	Method 1	Method 2
0	--	--
5	1.4	1.0
10	21.5	49.2
15	82.8	35.0
Bottom	--	--

6. SUMMARY AND CONCLUSIONS

The Columbia River Estuary has proved to be both qualitatively and quantitatively different from other estuaries for which conclusive studies of the salt balance have been made. The estuary is characterized by large hydraulic forces resulting primarily from the interaction of great river discharge and strong tidal currents. Large vertical salinity gradients can exist in the estuary as is expected for rivers with large freshwater flow. Though it sometimes meets the criteria for classification as a salt wedge estuary, the estuary is subjected to strong tidal influence which provides sufficient vertical mixing to make the two-layer approximation a poor one. In fact, salt is usually extended over the full depth of the estuary as a result of the tidally caused vertical mixing.

The estuary is quite short relative to its width. The distance of the salinity intrusion is controlled primarily by the tidal stage and is of the same order as the tidal excursion. Because the estuary is short, the salinity gradient is large in the horizontal as well as in the vertical. Despite the large density difference associated with the short transition region from river to oceanic conditions and the vertical stratification, the density-current structure is weak. Reversal of the net flow does not take place as near as 8 km from the mouth although manifestations of the density current are visible in the distribution of the nontidal drift.

Due to the weakness of the density-current strucuture, the longitudinal salt balance cannot be maintained primarily through the advection of salt

by the nontidal circulation. The analysis of the processes maintaining the salt balance during the high discharge stage indicates that the salt advected seaward with the river flow is balanced primarily by an upstream flux of salt associated with covariance of tidal-period fluctuations in current and salinity and by advective shear effect. This is a different result than is expected for the salt balance relation in a stratified estuary. The turbulent transport of salt by short-period fluctuations is usually large enough to be of at least secondary importance while contributions from other terms of tidal period are small.

The shear effect term arises in estuaries primarily from density currents and from velocity shear induced in tidal currents by bottom friction. Upon separation of the advective processes from the tidal processes in the shear effect term, the flux due to the density current is found to compose one-half to three-fourths of the total flux of the term. The harmonic expression derived to represent the shear effect term indicated that it could adequately be represented by a constant term plus two terms of tidal period. The constant term is interpreted as representing the density current while the terms with periodic variation are interpreted as the effects of tidal current shear since their maxima are near the maxima of tidal current. Even though the net flux of salt over a tidal cycle attributed to the alternating part of the shear effect can often be neglected, the tidal current shear can be important in the transport of salt for intervals of less than a tidal period because of the large vertical salinity gradient. For example, current reversal from ebb to flood takes place first near the bottom of the channel where bottom friction is important allowing high salinity water to be advected landward at depth while the

mean flow is still seaward. The stage of the tide is therefore a controlling factor in determining the dominance of the tidal current shear or density current in the shear effect term at any given time.

Computation of the fluxes and their corss-channel variations show a tendency for stronger flood current conditions to exist on the north side of the estuary coupled with stronger ebb conditions on the south side. Because of the difference in flood and ebb strengths across the estuary, the maximum salinity intrusion takes place in the northern channel, and its extent and duration are dependent upon the tidal stage. Neal (1965) stated that this cross-channel flow difference must be taken into account in the positioning of an outfall emitting a pollutant as this would have an influence on the resultant distribution of the added pollutant.

The accelerations related to channel shape have been shown to have at times greater influence on the dynamics of the estuary than does the Coriolis effect. These accelerations account for the lateral variations which were found. They will, however, complicate studies of the dynamics of the estuary. Current reversal from flood to ebb occurs first on the northern side of the channel at the Cape Disappointment section. This reversal takes place first on the southern side at the Clatsop Spit section which is also accounted for by the accelerations related to the shape of the channel. The tendency for reversal to occur first either at the surface or at the bottom, depending upon the tidal stage considered, conforms to the pattern for typical estuarine circulation.

The effective coefficient of horizontal eddy diffusion is extremely large when compared with similar values estimated for other estuaries by other workers. The increased size of this coefficient is indicative of

the large and readily recognizable role of tidal diffusion in the longitudinal salt balance. The comparability of the values estimated for the effective coefficient of vertical eddy diffusivity to values found in other estuaries indicates that, despite the strong tidal currents, gravity limits the effects of vertical mixing to that found in these other estuaries.

Similar conclusions for the period of low discharge were reached by Hansen (1965).

BIBLIOGRAPHY

Arons, A. B., and H. Stommel

1951. A mixing-length theory of tidal flushing. Trans. Am. Geophys. Un., 32(3):419-421.

Bowden, K. F.

1963. The mixing processes in a tidal estuary. Int. J. Air Wat. Pollut., 7(4-5):343-356.

Budinger, T. F., L. L. Coachman, and C. A. Barnes

1964. Columbia River effluent in the northeast Pacific Ocean, 1-61, 1962: selected aspects of physical oceanography. Univ. of Washington, Dept. of Oceanography Tech. Rept. No. 99, 78 pp.

Burt, W. V.

1956. Hydrography of Oregon estuaries prior to June (1956). Oregon State College School of Science. Ref. 56-2, 22 pp.

Burt, W. V., and B. McAlister

1958. Hydrography of Oregon estuaries, June 1956 to September 1958. Oregon State College School of Science, Data Report 3. Ref. 58-6, 18 pp.

1959. Recent studies in the hydorgraphy of Oregon estuaries. Res. Briefs Fish Commn. Ore., 7(1):14-27.

Hansen, D. V.

1964. Similarity solutions for salt balance and circulation in partially mixed estuaries. Ph.D. dissertation, Univ. of Washington, Seattle, 76 pp.

1965. Currents and mixing in the Columbia River estuary. Ocean Science and Ocean Engineering 1965, Vol. 2, The Marine Technology Society, Washington, D. C., pp. 943-955.

Hansen, D. V., and M. Rattray, Jr.

1965. Gravitational circulation in straits and estuaries. J. Mar. Res., 23(2):104-122.

Hickson, R. E., and R. W. Rodolf

1951. History of Columbia River jetties. Proc. 1st Conf. Coastal Engineering, Long Beach, 1950, ch. 32:283-298.

Kent, R. E.

1958. Turbulent diffusion in a sectionally homogeneous estuary. Johns Hopkins Univ. Chesapeake Bay Inst., Tech. Rept. XVI, Ref. 58-1: 86 pp.

Ketchum, B

1951. The exchanges of fresh and salt water in tidal estuaries. J. Mar. Res., 10(1):18-38.

1955. Distribution of coliform bacteria and other pollutants in tidal estuaries. Sewage and Industr. Wastes, 27:1288-1296.

Lockett, J. B.

1963. Phenomena affecting improvement of the lower Columbia estuary and entrance. Proc. 8th Conf. Coastal Engineering, Mexico City, 1962, ch. 40:695-755.

McAlister, W. B., M. Rattray, Jr., and C. A. Barnes

1959. The dynamics of a fiord estuary: Silver Bay, Alaska. Univ. of Washington, Dept. of Oceanography Tech. Rept. No. 62, 70 pp.

Neal, V. T.

1965. A calculation of flushing times and pollution distribution for the Columbia River estuary. Ph.D. dissertation, Oregon State Univ., Corvallis, 82 pp.

O'Brien, M. P.

1952. Salinity currents in estuaries. Trans. Am. Geophys. Un., 33(4):520-522.

Pritchard, D. W.

1952. Estuarine hydrography. Advances in Geophysics, Vol. 1, Academic Press Inc., New York, pp. 243-280.

1954. A study of the salt balance in a coastal plain estuary. J. Mar. Res., 13(1):133-144.

1955. Estuarine circulation patterns. Proc. Am. Soc. Civil Engineers. 81, Separate No. 717, 11 pp.

1958. The equations of mass continuity and salt continuity in estuaries. J. Mar. Res., 17:412-423.

Stewart, R. W.

1957. A note on the dynamic balance in estuarine circulation. J. Mar. Res., 16(1):34-39.

Stommel, H.

1953. Computation of pollution in a vertically mixed estuary. Sewage and Industr. Wastes, 25(9):1065-1071.

Stommel, H. and H. G. Farmer

1952. On the nature of estuarine circulation. Woods Hole Oceanographic Inst., Tech. Rept. Ref. 52-63.

U. S. Army Corps of Engineers

1960. Interim report on 1959 current measurement program, Columbia River at mouth, Oregon and Washington, Vols. I-IV, U. S. Army Engineer District, Portland, Oregon.

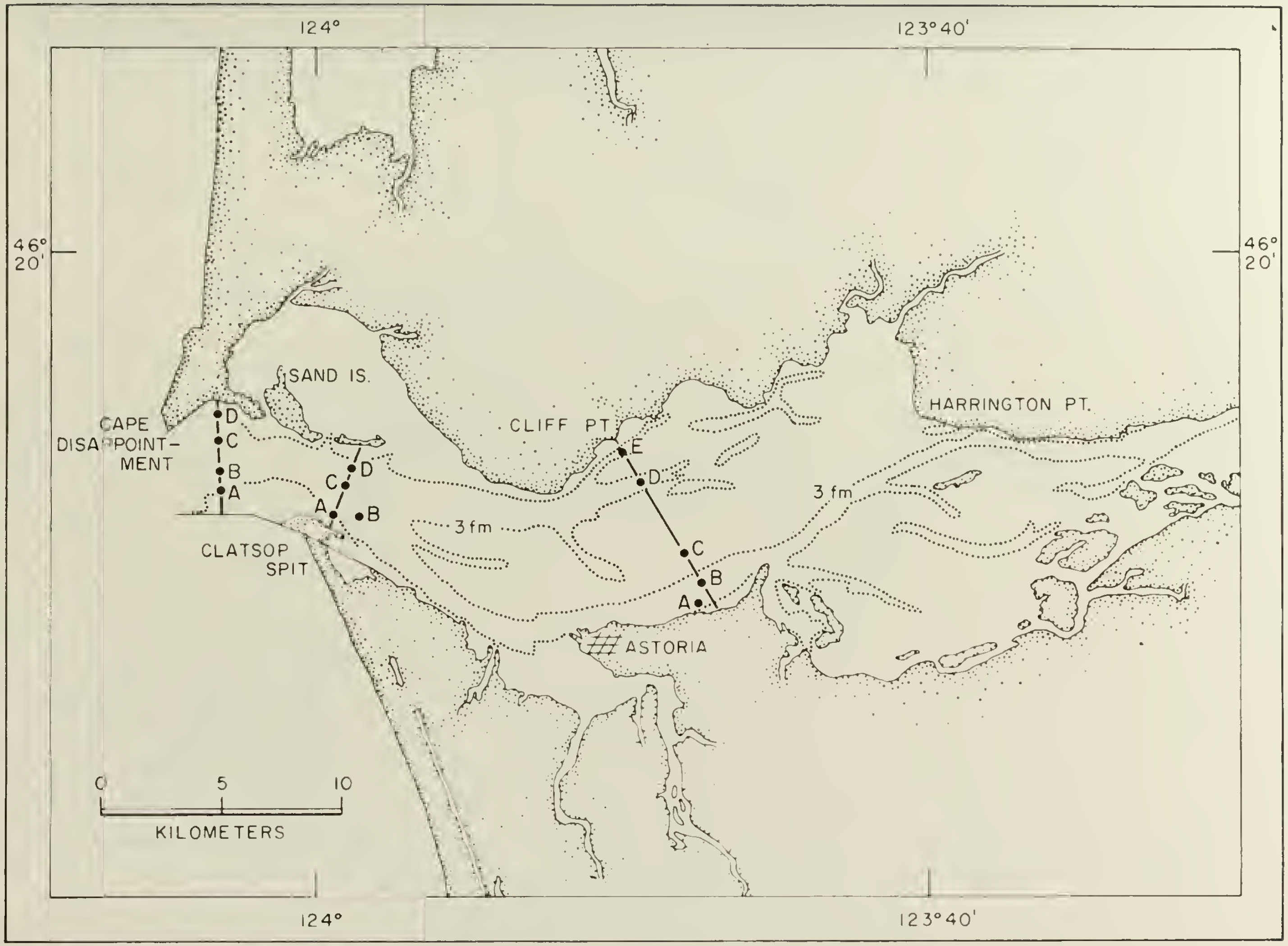

Figure 1. Columbia River Estuary showing station locations.

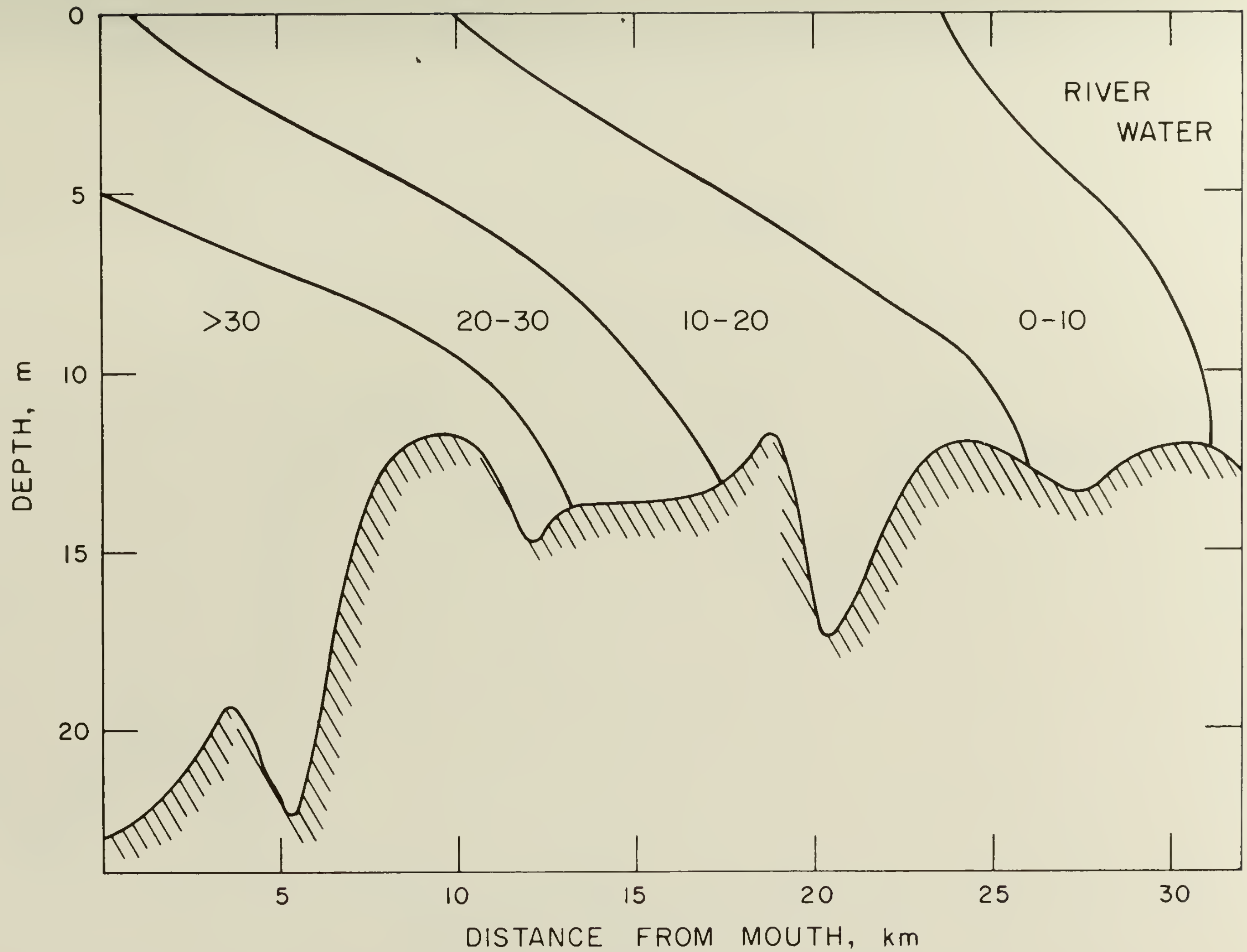

Figure 2. Longitudinal salinity distribution in the northern channel during higher high water and high river stage.

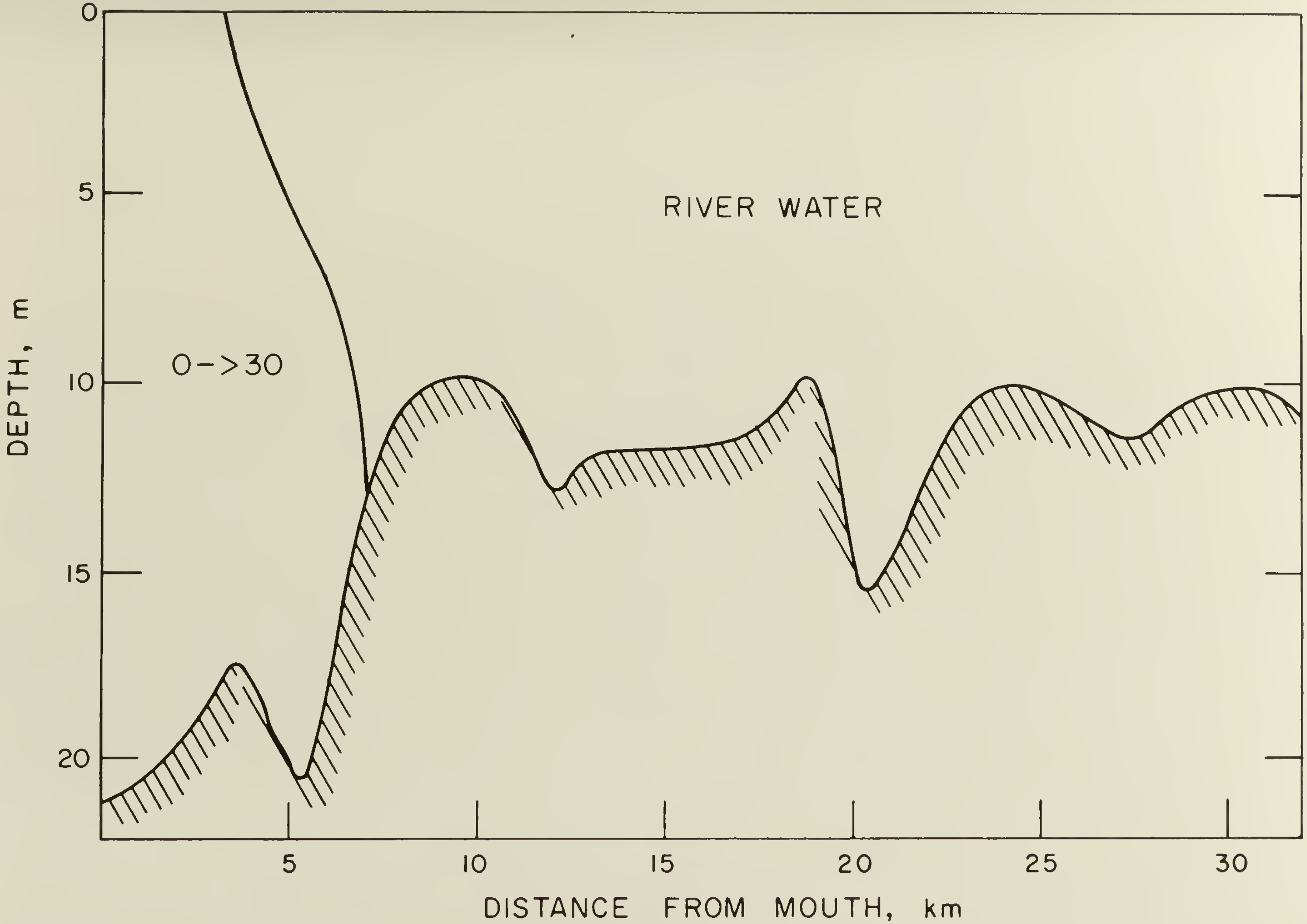

Figure 3. Longitudinal salinity distribution in the northern channel during lower low water and high river stage.

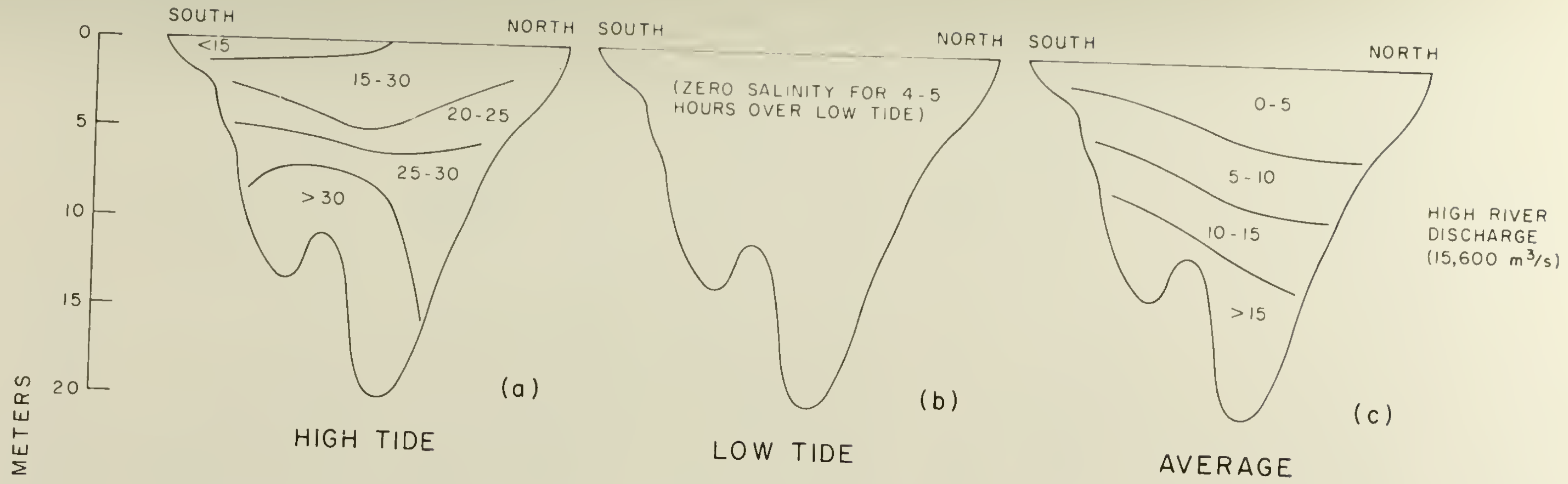

Figure 4. Salinity in per mille, Clatsop Spit section.

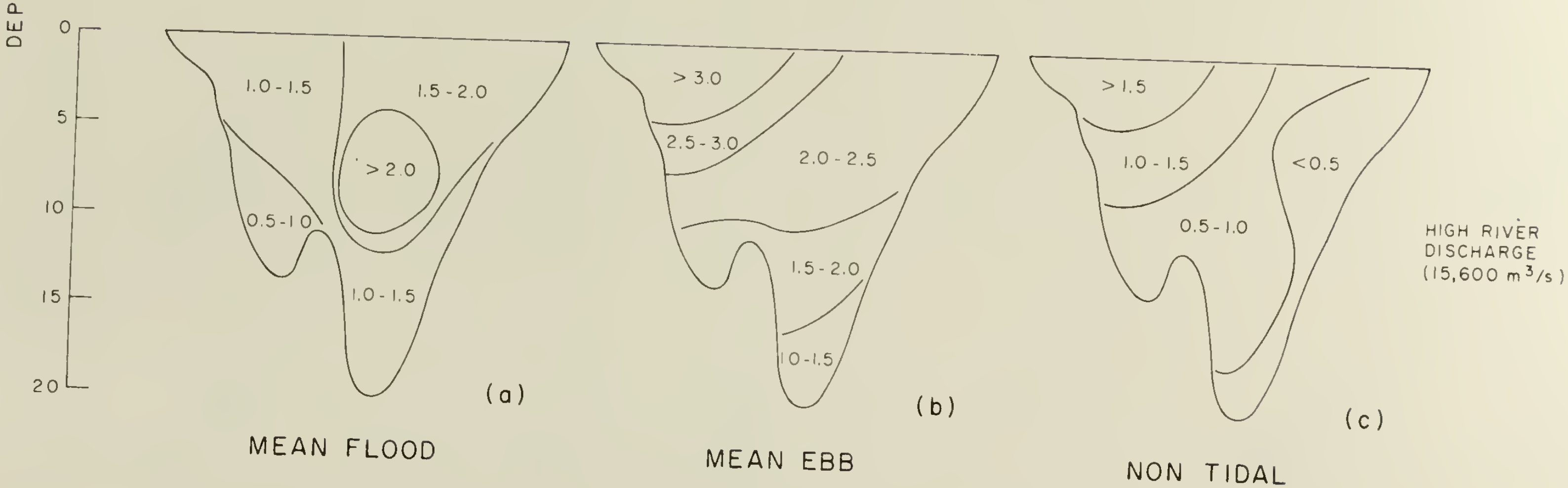

Figure 5. Current speed in knots, Clatsop Spit section (positive nontidal current in seaward direction).

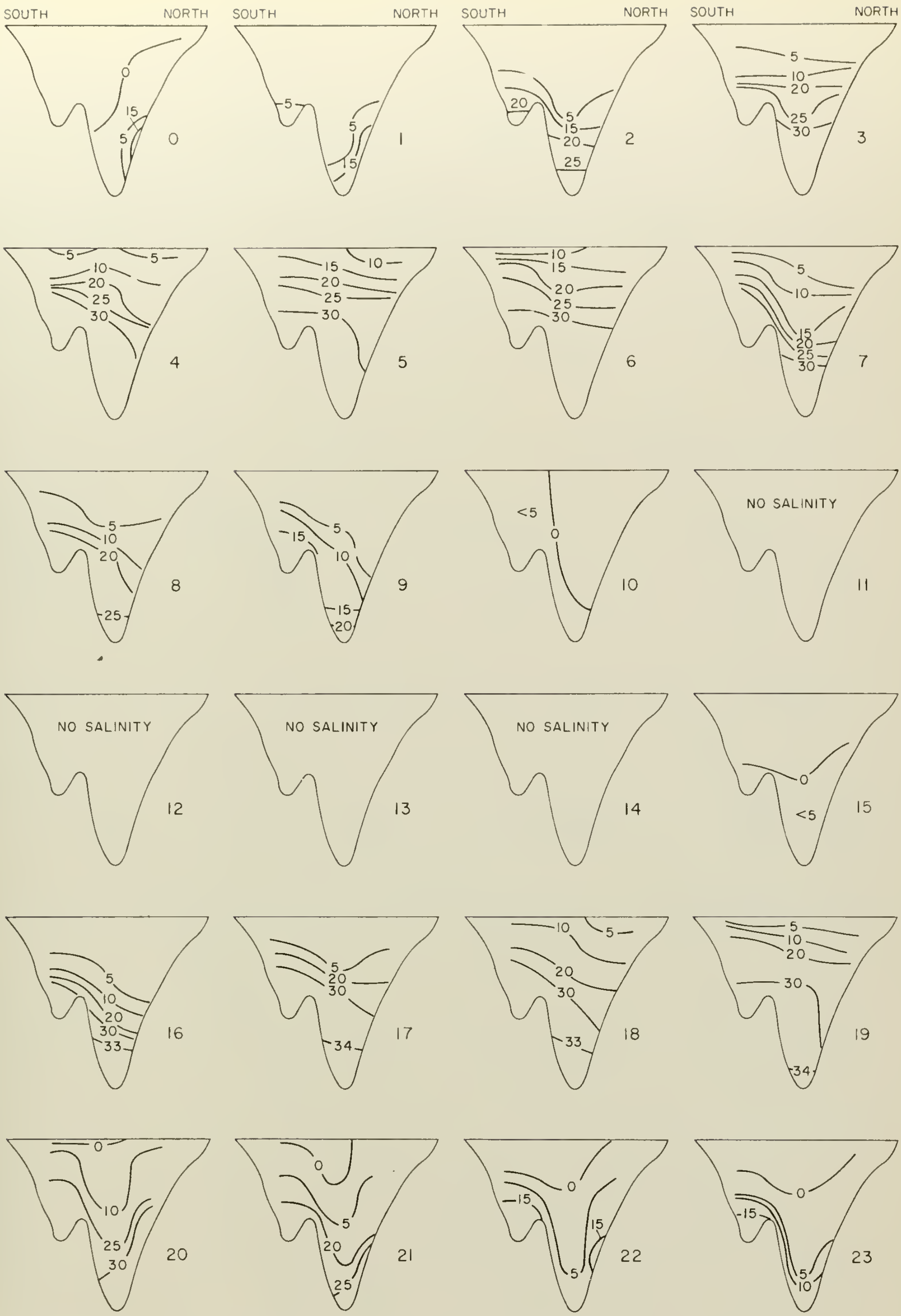

Figure 6. Variation of salinity over tidal period in increments of lunar hours, Clatsop Spit section. Salinity in per mille.

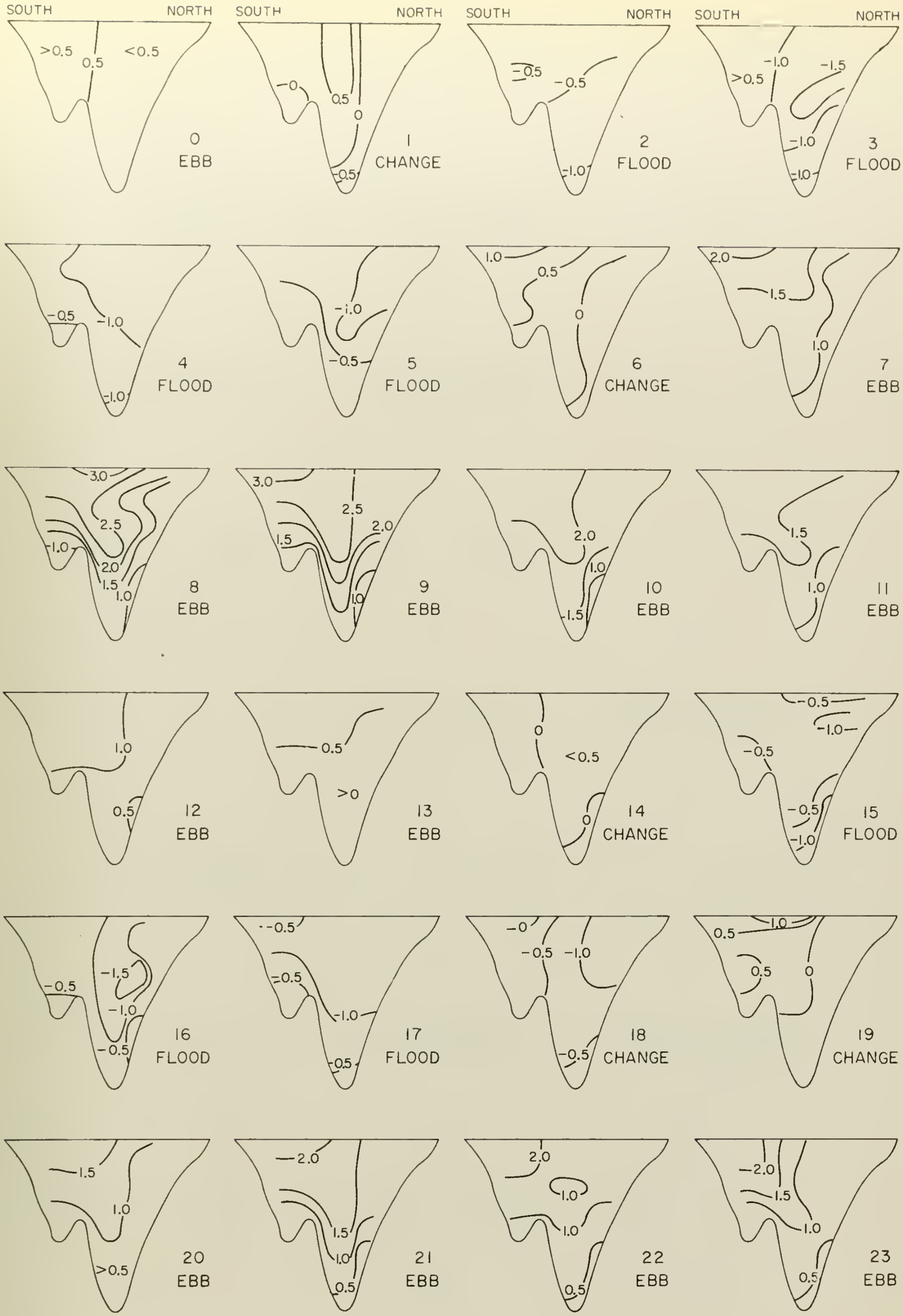

Figure 7. Variation of current speed over tidal period in increments of lunar hours, Clatsop Spit section. Current speed in m/s.

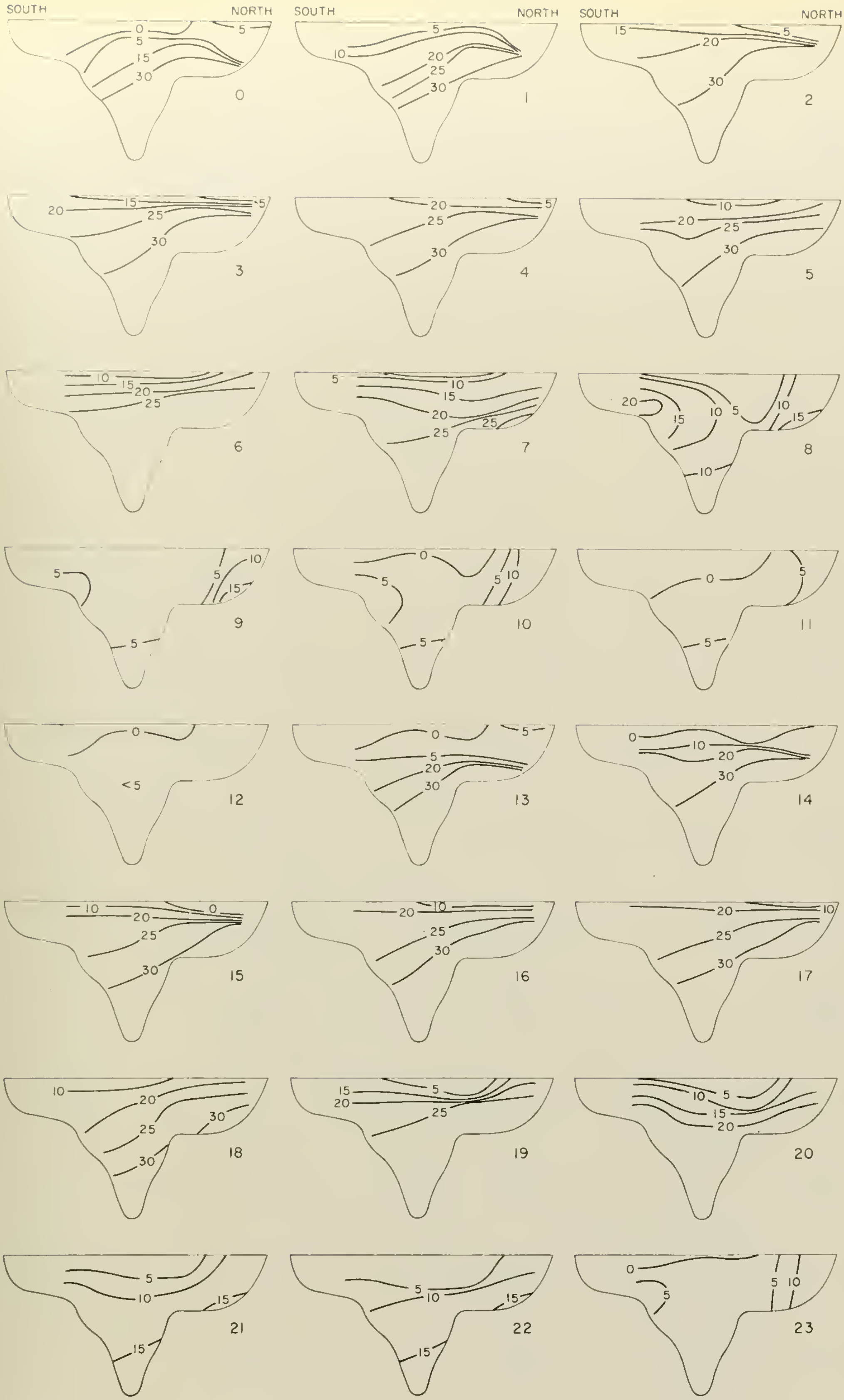

Figure 8. Variation of salinity over tidal period in increments of lunar hours, Cape Disappointment section. Salinity in per mille.

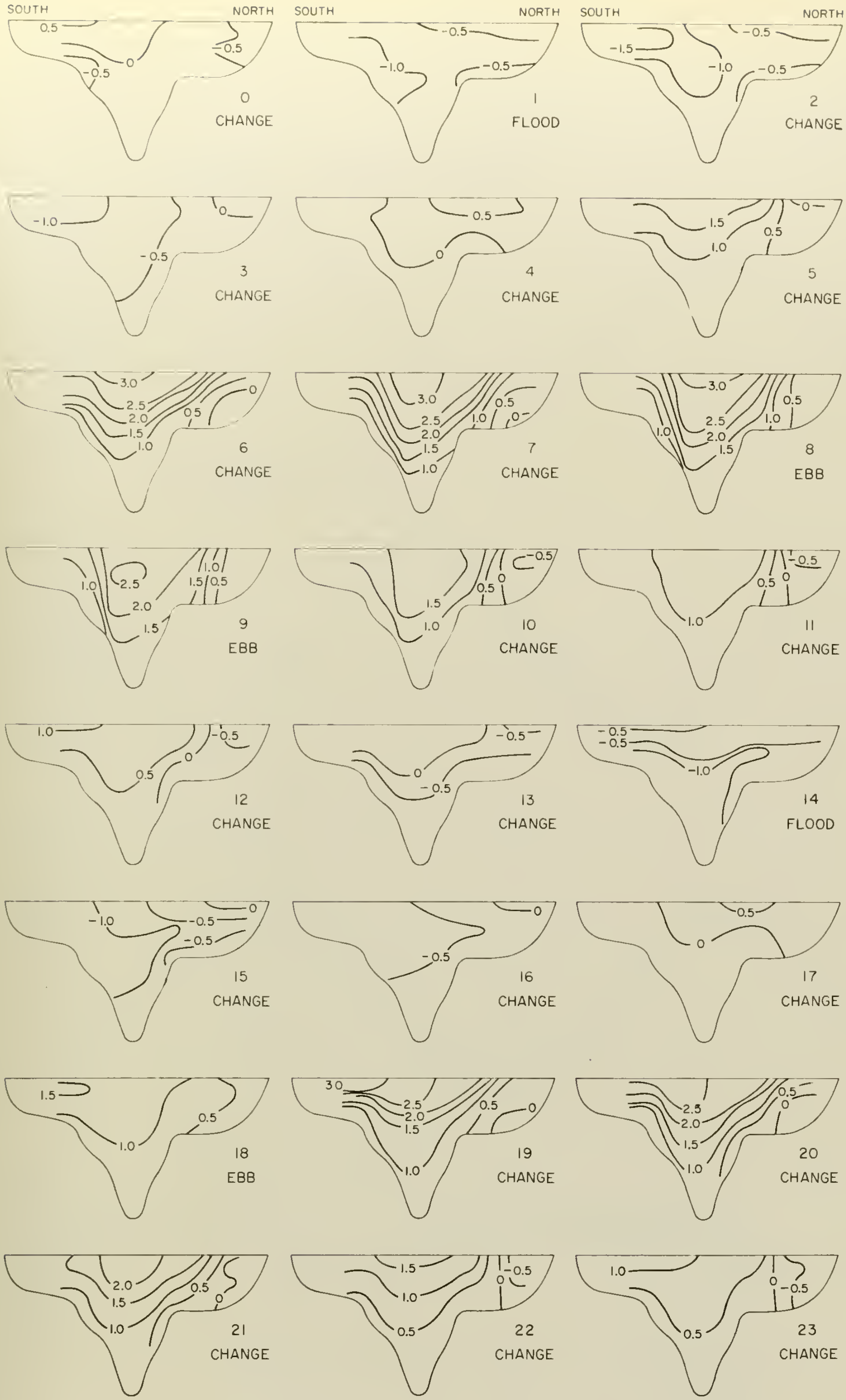

Figure 9. Variation of current speed over tidal period in increments of lunar hours, Cape Disappointment section. Current speed in m/s.

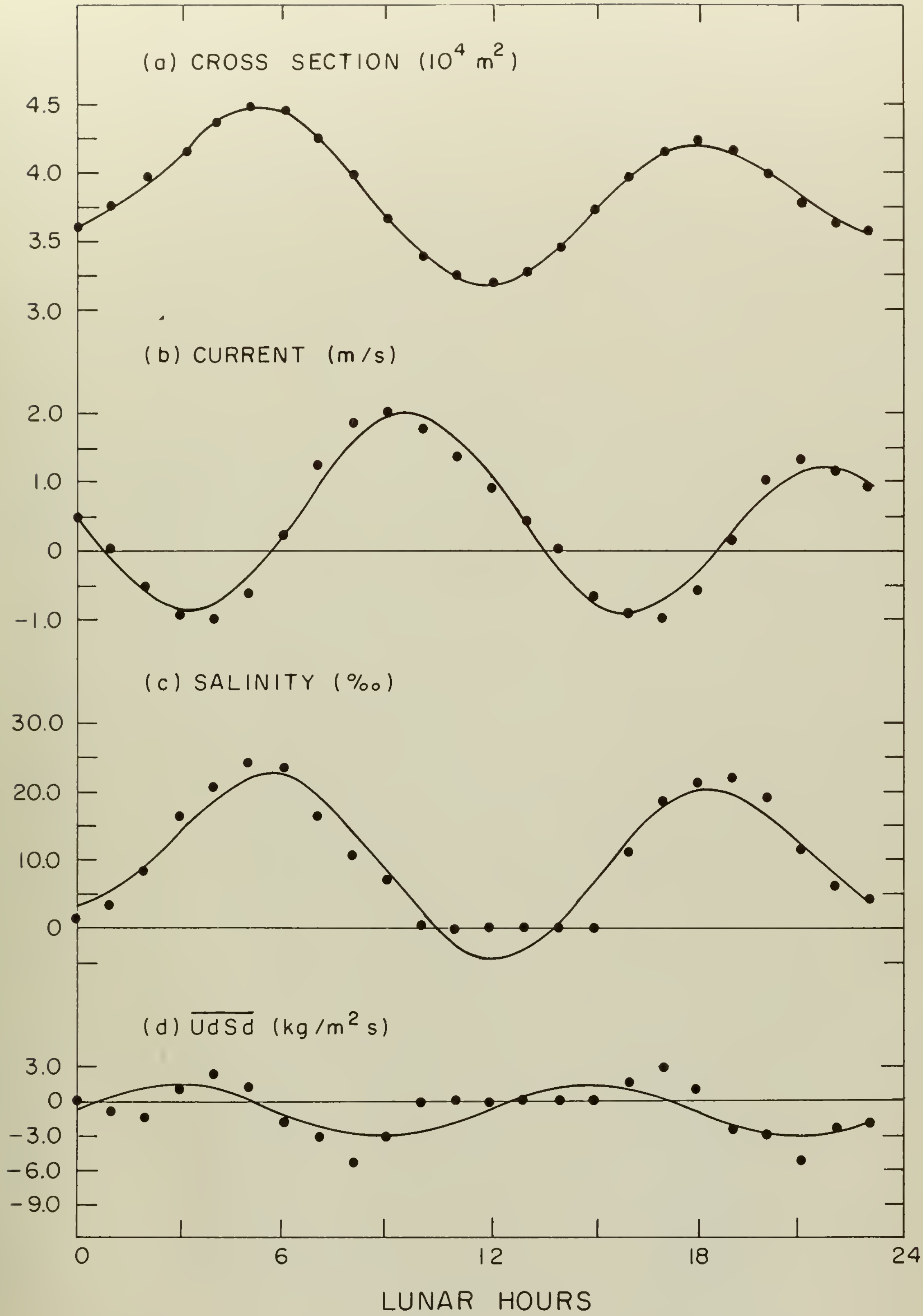

Figure 10. Tidal variations of mean conditions at high river stage, Clatsop Spit section.